TIBET UPLIFT AND EROSION

Eric J. Fielding[‡]

ABSTRACT

The 5-km-high Tibetan plateau is an outstanding topographic feature on the earth today. Its areal extent, elevation, and location cause significant effects on modern atmospheric circulation anti climate, so the history of uplift of the surface of Tibet is likely linked to Cenozoic climate changes, at local, regional, and perhaps global scales. Geological and geophysical studies of the plateau are contributing data on the present and past deformation of the Tibetan lithosphere that has formed the plateau, primarily during the Cenozoic. The principal of isostasy then can be used to estimate the elevation history of the surface for a given deformation history. Early Cenozoic north-south distributed shortening of Tibetan crust and mantle lithosphere probably caused significant uplift of the surface relative to sea level to perhaps half of the present elevation by the early Miocene. Thinning of the high-density mantle portion of the lithosphere during the Miocene may then have allowed the thick Tibetan crust to rise close to its present elevation (perhaps higher) before ~14 Ma, Since then, slow east-west extension of Tibet probably reduced the crustal thickness slightly and may have caused the elevation of the plateau to decrease during the late Cenozoic.

Erosion of Tibet, unlike narrow mountain belts, has been unable to match uplift of a broad plateau. Orographic precipitation and efficient river networks concentrate erosion on the edges, while the interior is protected from significant erosion despite its lofty elevation. The southern edge of the plateau, the

‡ Jet propulsion Laboratory, California Institute of Technology, Mail Stop 300-233, 4800 Oak Grove Drive, Pasadena, California 91109, USA

Himalaya, has suffered a minimum of 25 km of denudation since the Miocene, while central Tibet shows little or no sign of major erosion during that time. The Gangdese arc in southern Tibet apparently was rapidly eroded during the mid-Miocene when >4 km of rock were removed from the surface, as shown by mineral cooling ages. This pulse of erosion was probably caused by a combination of local thrust-system movement and changes in base level and precipitation due to relative elevation changes between the Gangdese and the Himalaya to the south, The modern long-wavelength flatness of Tibet is unlikely to have been caused by erosion and indicates viscous flow at some level of the lithosphere has been acting to level the plateau surface.

7. Introduction: Largest terrestrial plateau

Tibet is the largest and highest plateau on earth, a continuously elevated area nearly 3500 km by 1500 km in extent, It has resulted from the convergence of the Indian and Asian continental plates (Figure 1; Argand, 1924; Gansser, 1964; Powell & Conaghan, 1973; Molnar & Tapponnier, 1975). The plateau includes 82% of the world's land surface area that is >4 km above sea level and has an average elevation (for the internally drained portion) of 5023 m above sea level (Fielding *et al.*, 1994). The vast volume of the Tibetan Plateau is much larger than all the other presently active erogenic belts on the earth (such as the Altiplano-Puns plateau of South America, the Southern Alps of New Zealand, and Taiwan), as can be seen in the cross-sections of Figure 2. The high topography of Tibet strongly influences climate by affecting atmospheric circulation, not only locally but also regionally and perhaps over the entire northern hemisphere (Ruddiman & Kutzbach, 1989; Ruddiman *et* a/., 1989; Kutzbach *et al.*, 1993). The plateau is arid to semi-arid over most of its area and central Tibet is internally drained (Figure 1). Conversely, the edges of Tibet, especially on the south in the Himalaya, are wet, externally drained, and the site of intense erosion that may have caused a decrease of atmospheric carbon dioxide and global cooling (*e.g.,* Raymo & Ruddiman, 1992),

Several excellent reviews of the geological, geophysical and thermo-chronometric constraints on the various proposed mechanisms and consequences of the uplift of Tibet have appeared in the last decade (e.g., Dewey *et al, 1988;* Molnar, 1988; Burchfiel & Royden, 1991; Harrison *et a/.,* 1992a; Molnar et *al.,* 1993). This review intends to cover newer developments on the subject, present some different interpretations, and distinguish between the evidence for bedrock uplift and denudation histories.

1.1 Tectonic models

A wide variety of mechanisms for the uplift of the Tibetan Plateau have been proposed, extending back at least 70 years to Argand (1924). The many competing hypotheses for the origin and evolution of Tibet have been reviewed numerous times (e.g., Powell & Conaghan, 1973; Dewey *et al.,* 1988; Molnar, 1988; Burchfiel & Royden, 1991; Harrison *et al.,* 1992; Molnar *et a/.,* 1993), but I'll summarize some of the main points here. Most of the hypotheses attempt to answer three basic tectonic questions about the plateau: 1) how did the crust of Tibet become 50–75 km thick, twice the normal thickness of continental crust (e.g., Molnar, 1988; Him, 1988),2) what mechanisms absorbed the --2,500 km of convergence between the Indian and Asian plates since 40–50 Ma (Patriat & Achache, 1984; Besse & Courtillot, 1988; Dewey et a/., 1989; Le Pichon *et* a/., 1992), and 3) when did the plateau surface reach its present 5 km elevation above sea level. Clearly, these three questions are related and a complete Tibet uplift theory should answer all three. After briefly reviewing the tectonic mechanisms proposed to answer the first two questions, this paper will concentrate on the third question.

Some hypotheses propose that the double thickness of crust has formed by continual north-south shortening of the Asian crustal column (e.g., Dewey & Burke, 1973; England & Houseman, 1986), while other hypotheses posit that the double crust has formed by underthrusting of the Indian crust beneath Tibet (e.g., Argand, 1924; Powell & Conaghan, 1973; Barazangi & Ni, 1982). Both of these mechanisms would absorb a major (perhaps 1000 km or more) portion of the convergence between India and Asia within or beneath Tibet. Another set of hypotheses suggests that a significant part of the convergence between India and Asia was accommodated by motion on strike-slip faults that extruded blocks laterally (Molnar & l-apponnier, 1975; Tapponnier *et al.*, 1986; Peltzer & Tapponnier, 1988). This mechanism does not contribute to the thickening of Tibetan crust. Early versions of all of these hypotheses assumed that the given mechanism operated reasonably continuously since the beginning of the collision, but many variations in timing and duration have been proposed, as described below. Another mechanism, first modeled by Houseman *et al.* (1981), can cause uplift of the plateau without changing the crustal thickness. Houseman *et al.* (1981) and England & Houseman (1988; 1989) propose that part of the lithospheric mantle of Tibet has been removed and replaced with ıotter, less dense asthenosphere. This would have the effect of decreasing the mass of the lithospheric column and could result in isostatic uplift of 1–3 km.

Many recent theories propose that different mechanisms have operated at various times since the first contact of the Indian and Asian continental plates. The precise date of the final closure of the Neotethys ocean formerly separating India and Asia is still uncertain (and probably was somewhat diachronous along the nearly 3000 km long margin), but dates in the Paleocene–Eocene near 45–55 Ma are becoming widely accepted (e.g., Patriat & Achache, 1984; Searle et *al.*, 1987; Besse & Courtillot, 1988; Dewey *et al., 1989).* Much of the difference between times for the initial "collision" is due to various definitions of the term, and the exact timing is relatively unimportant to the discussion in this paper, so I will arbitrarily use ~50 Ma. The current increase in data on both the structure and history of Tibet and adjoining areas requires more complex theories to match all the newly available facts, In particular, there is now strong evidence that certain structures were active for subsets of the time since -50 Ma, which suggests that the deformation mechanisms expressed by those structures were similarly restricted in time (e.g., Harrison *et a/.,* 1992a).

Tibet was closed to foreigners after the 1920's so the only westerners to do extensive geological work, until the early 1980's, were those of the Sven Hedin expeditions of the early part of this century (e.g., Hennig, 1915; Norin, 1946) and the Littledale expedition of the late 1800's (Littledale, 1896). The Hedin expeditions provided one firm constraint on the uplift history by reporting the widespread shallow-water marine Albian limestones deposited in southern Tibet (Hennig, 1915; Norin, 1946). This indicates that at least those parts of southern Tibet were at, or near, sea level during the mid-Cretaceous. In the past decade or two, detailed field studies have been conducted in only a few scattered locations, primarily around the edges of the plateau and near Lhasa. This means most available chronological information on the post-C setaceous history is concentrated near or beyond the margins of the Tibet, The constraints on the uplift and erosion histories will be discussed in more detail below.

While there is still considerable controversy about the differing tectonic models for the formation of Tibet, some consensus is beginning to emerge about certain aspects. Paleomagnetic measurements and plate reconstructions, while not precise, show that the Indian Plate has moved 2600±900 km northward relative to the Eurasian Plate since the closure of the Neotethys (Patriat & Achache, 1984; Besse & Courtillot, 1988; Dewey *et a/., 1989;* Le Pichon et *a/.,* 1992). Paleomagnetic samples from Tibetan rocks near Lhasa indicate a northward displacement of 20°±6° (roughly 2000 km) for southern Tibet relative to stable Eurasia (Achache *et aL,* 1984; Besse & Courtillot, 1988), but the uncertainties in this measurement are large so it does not provide a strong constraint on tectonic models.

Several studies using satellite and seismic datasets have contributed to our understanding of the neotectonics and present structure of the crust and upper mantle for the vast interior of Tibet and surrounding areas. Earthquake sources provide information on the modern deformation of the crust (and mantle), while the propagation of the seismic waves provides data on the present velocity and attenuation structure of the crust and mantle through which they pass. Molnar (1988) reviews the many seismic constraints for Tibet. Soon after the launch of the first Landsat (initially called ERTS) satellite, Molnar & Tapponnier (1975) used Landsat images to produce a new tectonic interpretation and map of the far-flung effects of the India-Asia collision. Since then, many studies have used satellite imagery to interpret tectonic features in Tibet, many by supplementing the attributes mapped in the field (Ni & York, 1978; Molnar & Tapponnier, 1978; Rothery & Drury, 1984; Armijo e? a/., 1986; Zhou *et al.,* 1986; Kidd *et al.,* 1988; Kidd & Molnar, 1988), The present tectonic regime of the Tibetan plateau clearly involves some component of east-west extension as shown by earthquake focal mechanisms (e. g., Molnar, 1988) and field mapping and interpretation on airphotos and Landsat imagery of normal and strike-slip faults (Armijo *et a/., 1986;* 1989; Kidd & Molnar, 1988).

1.2 **Plateau morphology**

Fielding *et* a/. (1994) presented analyses that quantify the modern topographic characteristics of Tibet and surrounding areas, especially the extraordinary flatness of the plateau at a wide range of scales, I summarize the results here. The Tibetan Plateau has abrupt boundaries to the north and south (Figure

3). Tibet is bounded on the south by the high and rough Himalayan ranges and on the north by the Kun Lun mountains. To the east and west, the high topography extends beyond the political boundaries of Tibet, The western part of Tibet is continuous with the Karakoram and Pamir mountains, which are considered here as part of the plateau, and the eastern part of Tibet slopes downward until it reaches the Longmen Shari at the edge of the Szechwan basin.

The central and northern parts of the plateau have the lowest relief, on both short and long wavelengths, The dark shading on Figure 4 shows the vast areas with low slopes of less than 10° (Fielding *et a/., 1994).* The slopes steepen only near a few glaciated peaks and near the Pliocene-Quaternary grabens in the south and south-central part of the plateau (e. g., Armijo et *al.*, 1986). Average slopes over -250 m wavelengths are about 5° in central Tibet. At moderate wavelengths of the 100 km wide swaths of Figure 5, relief is about 1 km or less for most of Tibet, as opposed to the much higher relief of up to 6 km on the plateau edges (Fielding *et al.*, 1994),

At the longest wavelengths, the hypsometry demonstrates that more than half of the internally drained area lies within ±200 m of the mean elevation (Figure 6) of 5023 m above sea level (plotted on Figure 6; Fielding *et a/.,* 1994). The total area is 5,5 x 10^5 km^2. The modal elevation (between 4920 and 4960 m), where the area within a single 10 m elevation bin reaches 1.01 x 10^4 km^2 or 1.8°/0 of the total area, is slightly less than the mean indicating a small degree of skew, although surprisingly close to a symmetric distribution.

2. ***Uplift history***

In Tibet, as in most places, chronological data become sparser and generally less accurate the farther back one looks into its past. This is especially true when we consider data that constrain the paleoelevation of the plateau surface.

Paleobotanical determinations of uplift are controversial and subject to confusion with climate changes (Molnar & England, 1990; Molnar *et al.*, 1993) and will not be discussed here. Some geologic dates constrain deformation history that can be related to surface elevation through isostasy. Most geological information, such as mineral cooling ages, is more directly related to erosion than uplift of the surface (England & Molnar, 1990). Erosion history will be discussed in the next section (§ 3).

England and Molnar (1990) reviewed many of the problems with determining uplift of the surface of a mountain belt with respect to a stable reference, such as the geoid. The geoid itself is deformed slightly by the redistribution of mass within the lithosphere, such as forming a mountain range, but the largest variations in the elevation of the geoid with respect to a spheroid or datum (up to ~100 m; Lerch *et al., 1994)* are associated with lateral density variations deep in the mantle and perhaps at the core-mantle boundary (*e.g.,* Bowin, 1991). The geoid is an equipotential surface so it defines the extension of "sea level" over land areas. However, changes in the geoid elevation over geologic time, due to mantle convection, are likely to be nearly as large as the eustatic sea-level changes (and geoid changes can be confused with eustatic changes in the geologic record), so the geoid is not a long-term stable reference either, In any case, the variations of the geoid and eustatic sea-level elevations in the past are less than a few hundred meters, which is much less than the errors on paleoelevation estimates, so the exact reference choice is unimportant to this study. I will use the term "sea level" below.

It is convenient to break up the history of Tibet uplift into several stages for the discussion below. The timing of the transitions between each of these stages is still controversial, but at least three of the four phases below are well-accepted as separate stages of the deformation history: the pre-collision phase, the collision-induced crustal thickening phase, and the late extension phase. I distinguish between the Early Tertiary collision and Middle Tertiary compression phases because of the growing number of indications that the deformation pattern changed dramatically sometime around 25–30 Ma (e.g., Harrison *et al.,* 1992a). For each phase, I attempt to estimate the elevation of the plateau that would result from isostatic balance of the crustal and lithospheric thickness predicted by various tectonic models (see Fig. 7) in section §4 below,

2.1 Recent

The only precise constraint on the uplift history of the Tibetan plateau is the present elevation of 5.02 km relative to the WGS84 geodetic datum for the central plateau (Fielding *et al.,* 1994). The various uplift histories of different parts of the plateau must converge on this elevation by the present to match the unusual level surface (Figs. 5 & 6). The surface elevation is determined by the isostatic balance of the density distribution versus depth. Because the plateau is so wide (> 600 km at the narrowest point), local isostasy can be assumed because flexural support is limited to much shorter wavelengths (c 200 km) for a young orogenic belt. Gravity measurements from the interior of Tibet have free-air and isostatic anomalies close to zero, in agreement with local isostatic equilibrium (Tang *et al.,* 1981; Zhou *et al.,* 1981; Molnar, 1988; Jin *et al.,* 1994). Without other information about paleoelevations, one can calculate isostatic elevations from estimates of the crust and upper mantle structure at a given time,

The present average surface elevation in eastern Tibet slopes downward gradually from the 5 km elevation of the internally drained central Tibet to about 4 km above sea level at the eastern end (Figure 5b). This decrease in elevation eastward is presumably due to the isostatic response of an eastward decrease in crustal thickness, The elevation decrease starts approximately at the internal-external drainage divide which is a few hundred km west of the eastern (Namche

Barwa) syntaxis of the Himalaya (Figures 1 and 5b). The uplifted volume of eastern Tibet that is east of the syntaxis likely represents material extruded eastward (Dewey *et a/.,* 1989; Le Pichon *et al.*, *1992),* and the long-wavelength surface gradient (-1 km in -1000 km) of eastern Tibet provides an eastward horizontal buoyancy force that continues to drive this extrusion (England, 1995),

2.2 **Mesozoic-Early Tertiary pre-collision**

The one widely distributed constraint, described above, on the former surface elevation of what is now Tibet is the Albian/Aptian limestone deposits that indicate that much of the surface was slightly below sea level at roughly 100 Ma (Hennig, 1915; Norin, 1946). We also have to make a correction for the eustatic sea-level change since the mid-Cretaceous, when global sea level was considerably higher (> 100 m) than today, so it is convenient to set the elevation at about -200 m for that time. This paleoelevation was most likely supported by "normal" continental crust, although Tibet is made up of several terranes that have been accreted onto Asia from the Late Paleozoic through the Middle Mesozoic (e.g., Dewey *et al.,* 1988), so the crust and lithosphere may have been thinner or thicker than that of a stable craton. One simplistic approximation of the surface elevation history for Tibet is gradual uplift from ---0.2 km at -100 Ma to 5 km at O Ma, or an average rate of -0.05 mm/yr = 0.05 km/Ma (Figure 7). This is a lower bound on the uplift rate of the surface with respect to sea level.

From the mid-Cretaceous until the early Eocene (120-40 Ma), northward subduction of the Neotethys oceanic crust formed an Andean arc on the southern margin of Asia, now exposed as the extensive Gangdese (Transhimalayan), Ladakh, Kohistan, and Karakoram batholiths of southern Tibet and the Karakoram (e. g., Harris *et al.*, 1988; Dewey *et a/.,* 1988). If the modern Andean Cordillera can be taken as a tectonic analog, then significant crustal thickening within the arc and some crustal thickening in the back-arc probably occurred in the Tibetan terranes. The Andean Cordillera varies greatly along its -3000 km length with different regimes of deformation associated with different subduction geometries (e.g., Jordan *et al.*, *1983).* One might expect a similar variation of deformation along the Gangdese-Karakoram arc, but present data are insufficient to document this.

Two segments of the Andean Cordillera presently have "flat-slab" subduction, where the Nazca plate is essentially horizontal for some distance beneath the South America plate (Jordan *et a/.,* 1983). In the southern flat-slab region, the lithosphere of the overriding continental plate has been thinned (scraped off?) to roughly 100 km thick, clearly shown by the earthquake depths in the Wadati - Benioff zone (Smalley & Isacks, 1987), Despite the relatively thick continental crust (40–50 km) and thin mantle lithosphere (45–55 km), average elevations over the flat-slab are held down to 1-2 km above sea level by the cold oceanic lithosphere below (Smalley & Isacks, 1987).

It's possible that at least part of Tibet suffered a similar episode of flat-slab or shallow subduction as suggested by Coulon *et a/.* (1986) and Copeland *et a/.* (1995). While sampling is not complete, there is an apparent gap between 90 and 70 Ma in the ages of Gangdese arc batholiths (Harris *et al.*, 1988) that could have been caused by a flat-slab. Dates have been obtained for an andesite with an $^{40}Ar/^{39}Ar$ age of 76.6±1.5 Ma and a dacite with a Rb-Sr whole-rock age of 80.5±1.4 Ma near Amdo in central Tibet, some 250 km behind the arc north of Lhasa (Coulon et al., 1986). Samples from other sites in the northern part of the Lhasa block also indicate widespread volcanism between 110 and 80 Ma (Coulon *et al.*, 1986), Both main-arc magmatic shut-off and magmatism far behind the arc are observed in the Andes over the flat-slab areas (Cross & Pilger, 1982; Jordan *et a/.,* 1983; Kay *et a/.,* 1987). If there were shallow or flat-slab subduction beneath Tibet, then the lithosphere would have been thinned and reduced in strength. This region of reduced lithospheric strength could then have shortened relatively quickly to produce an early Tibetan plateau, similar to the scenario proposed by Isacks (1988) for the formation of Altiplano-Puns plateau.

Burg *et a/,* (1983), Coulon *et al. (1986),* and Yin *et al. (1994)* report field evidence north of Lhasa and near Chuoqin -500 km to the west for a strong deformation event during the Cretaceus in the northern part of the Lhasa block, This deformation in the back-arc of the Gangdese arc in southern Tibet ended before the Paleocene deposition of the widespread Linzizong volcanic rocks. After that event, there was a lack of major shortening from the Paleocene to the late Oligocene–early Miocene within the Gangdese arc and back-arc (Lhasa block), despite substantial convergence of the Indian continental plate after its initial contact with the Lhasa block during this time (Burg eta/., 1983; *Yin et a/.,* 1994). England & Searle (1986) suggested that major crustal thickening of the Lhasa block, perhaps analogous to the Altiplano-Puns plateau in the central Andes, may explain both the evidence of shortening before the late Cretaceous–Paleocene and the lack of deformation after the continental collision. In such a scenario, there may have been a high (3-4 km), relatively narrow (-300 km) plateau (see Fig. 2 for the present profile of the Altiplano) in what is now southern Tibet with thick (60–70 km) crust and relatively thin lithosphere during the late Cretaceous–Paleocene, before the impingement of the Indian continent.

Several authors assumed that the elevation of the surface of Tibet above sea level just before the continental collision was 500 m or less (England & Houseman, 1986; Mercier *et a/.,* 1987; Dewey et *a/.,* 1989). Le Pichon *et al.* (1992) investigated the implications of different assumptions for the pre-collision elevation of Tibet and surrounding areas for the mass balance of crustal shortening, erosion, and extrusion. They concluded the lowest elevation assumption that Tibet and surrounds were at sea level is unrealistic, and they preferred a hypothesis that both Tibet and surrounding areas were 500 m above sea level before the contact of the Indian continent. A third postulate, that the Tibetan area was 1000 m and adjacent areas 500 m above sea level, was considered by Le Pichon *et al.* (1992) to be the highest elevation possibility and causes a 30 to 40% reduction of the amount of post-collision convergence that was converted to volume of the region above the pre-existing topography. The speculative Altiplano-Puns sized plateau described above would have had a pre-collision volume somewhat larger than the latter assumption analyzed by Le Pichon *eta/.* (1992) and require an even larger reduction of the post-collision convergence component of the volume.

2.3 **Early Tertiary collision**

The contact between Indian continental crust and the Asian plate in the Paleocene–Eocene (45–55 Ma) formed the Indus-Yarlung-Zangbo Suture (e.g., Gansser, 1964; 1981; Dewey *et al.,* 1988). Deformation of the northern margin of India began at that time (e.g., Searle, 1986; Searle *et a/.,* 1987), although the major crustal shortening along the Main Central Thrust (MCT) apparently began much later in the late Oligocene–early Miocene (Hubbard & Harrison, 1989). The complicated history of shortening of the Indian crust to form the Himalayas is outside the scope of this paper, but some 400–600 km (Ratschbacher *et a/.,* 1994) to >1000 km (Dewey *et a/.,* 1989) of convergence between India and Asia has been absorbed there. Other structures to the east of Tibet, such the Red River ductile shear zone, also absorbed some of the convergence by extruding parts of Tibet and China eastward along left-lateral strike-slip faults (Tapponnier *et al.,* 1982; Peltzer & Tapponnier, 1988), The Red River fault zone was active in a left-lateral sense from 35 Ma to 17 Ma with hundreds of km of shear (Harrison *et a/.,* 1992b; Leloup *et a/.,* 1993; Brias *et al.*, 1993). Some or most of the remaining convergence was taken up as crustal shortening within Tibet (e.g., Dewey *et a/.,* 1988; 1989; Le Pichon *et al.*, 1992).

The Royal Society -Academica Sinica team, in their reconnaissance field mapping along the Geotraverse route from Lhasa to Golmud in east-central Tibet, found clear evidence of shortening that affects redbeds in several locations (Kidd *et al.,* 1988). In most places, the age of the deformed strata was not well constrained, but the red elastics of the Fenghuoshan Group in the north-central part of the Geotraverse are Eocene (Kidd *et a/,, 1988),* A map and cross-section of the area near Erdaogou yield a minimum of 40% post-Eocene shortening by folding of the Fenghuoshan strata (Kidd *et a/.,* 1988; Coward et *a/.,* 1988), The lower bound on the age of deformation is somewhat constrained by a prominent thrust that places the Eocene redbeds over another marl and elastic unit that is much less lithified and presumably younger (Kidd et *al.*, 1988; Leeder *et al.*, 1988).

The apparent lack of shortening within the upper crust of southern Tibet (the Lhasa block) between the Paleocene and the late Oligocene–early Miocene (Burg *et a/., 1983;* Yin *et aL,* 1994) requires upper crustal shortening within Tibet during that time, if any, to be concentrated in the central and northern parts. This would be consistent with early folding of the Eocene Fenghuoshan strata in northern Tibet. England & Searle (1986) suggested that, if the crust of southern Tibet was already thickened substantially before the continental collision, due to the horizontal buoyancy force of the elevated southern Tibet block, little further shortening would occur there. Modeling with a thin viscous sheet showed dis-tributed crustal thickening (and uplift) propagating northward from southern Tibet to form of the rest of the Tibetan plateau as the Indian continent continued to move northward, This conjecture would produce an uplift history for central and northern Tibet similar to, but early than, other distributed shortening models (e.g., England & Houseman, 1989), while southern Tibet would be uplifted only moderately after the continental collision.

The transfer of convergence stress northeastward through Tibet to the Red River shear zone starting at -35 Ma would also be greatly enhanced by a substantially thickened crust and high elevations in Tibet (England & Houseman, 1986). Other mechanisms for thickening the crust of Tibet without pervasive deformation of the upper crust of the Lhasa block would include underthrusting of the Indian plate beneath that part of Tibet (Argand, 1924; Powell & Conaghan, 1973; Barazangi & Ni, 1982; Beghoul *et a/.,* 1993) and the upper crustal buckling-trough contraction model described by Burg *et al.* (1994). Detailed dating of the timing of deformation across Tibet would test nearly all of these hypotheses.

The elevation of the Tibetan surface above sea level that would be predicted by the various hypotheses for this stage differs greatly. The propagating distributed shortening and underthrusting models both predict a plateau of roughly double-thickness crust that grows progressively wider with time, but the thickness of the mantle lithosphere below will still affect the resulting elevations of the surface. Dewey *et al.* (1989) suggested that the crustal thickening in Tibet between the initial collision and ~30 Ma caused the surface elevation to reach 3 km, but did not predict the details of the resulting structure of the lithosphere. Unfortunately, there are no strong geologic constraints on the lithospheric structure at this time, so we have to work from assumptions.

As described above, there are several reasons to believe that the mantle lithosphere of Tibet was significantly thinner than an old cratonic lithosphere. The underthrusting model requires that the Tibetan mantle lithosphere be almost completely removed to allow the underthrust Indian crust to move directly under the Tibetan crust. The mantle lithosphere in the underthrust region would then be that of the Indian plate which is more cratonic in character. The distributed shortening model does not require any specific mantle lithosphere configuration, but the apparently rapid shortening of the Tibetan lithosphere would be facilitated by having a thin, weaker lithosphere to start with. After the distributed shortening, one would expect that the mantle lithosphere would have been shortened and thickened by the same factor as the overlying crust.

2.4 **Middle Tertiary compression**

Several well-dated events occurred in the late Oligocene–Miocene that indicate major changes in the mechanisms of deformation of Tibet and surrounding regions. Dewey *et a/,* (1989) surmised that north-south plane strain in Tibet changed at ~30 Ma to a wrench regime with some north-south shortening combined with east–west extension on conjugate strike-slip faults. In southern Tibet, Yin *et a/.* (1994) and Harrison *et a/.* (1992) document shortening on the Gangdese Thrust System (GTS) that thrusted the batholiths of the Gangdese arc southward over the IYZS starting at -27 Ma and continuing until -23 Ma. In some areas, the GTS was then overridden by a north-directed thrust, the Renbu - Zedong Thrust (RZT) that was active at 17.5±0.9 Ma (Ratschbacher *et a/.,* 1994; Yin *et a/.,* 1994). Major crustal shortening in the Himalayas, within the Indian plate in Nepal, along the Main Central Thrust (MCT) happened at roughly the same time: 21 Ma (Hubbard & Harrison, 1989) and before -22 Ma (Guillot *et a/.,* 1994). Apparently synchronous with at least some of the activity on the MCT, the South Tibetan Detachment (STD, also called the North Himalayan normal fault) locally extended and rapidly exhumed what is now the High Himalaya crystalline slab (Burchfiel *et* al., 1992). The extrusion of Indochina to the southeast along the Red River shear zone also seems to have begun at -35 Ma and ended at -17 Ma (Harrison *et a/., 1992b;* Leloup *et al., 1993).*

It is likely that at least the MCT, STD, and RZT movements are related since they are within ~200 km of each other and overlapped in time, and perhaps the

GTS was a precursor to the other three (Yin *et al.,* 1994). One possibility is that this surge of shortening in southernmost Tibet was caused by the thickening of the lithosphere in northern and central Tibet having exceeded a threshold where the horizontal buoyancy forces resisted further shortening there (England & Houseman, 1986; Dewey *et al.,* 1989). Burchfiel & Royden (1985) and Burchfiel *et al.* (1992) suggested that the movement on the MCT and STD represent a gravitational collapse of high topographic relief between southern Tibet and the Indian plains. Another possibility is that erosion of the southern margin of the plateau during the Miocene removed enough mass to cause the Indian continent to begin underthrusting Tibet, as proposed for the Altiplano-Puns by Isacks (1 988) and Gubbels *et al.* (1993). Whatever its cause, the deformation of southern Tibet represented by the GTS and RZT is a strong constraint on theories that predict crustal thickening progressing monotonically from south to north (and presumably progressive uplift of the surface as well).

Most of the hypotheses described above to explain the important change in the deformation regime of Tibet at 30-25 Ma also predict that the plateau had reached a reasonably high elevation by that time, at the beginning of this phase of the uplift history from -25 Ma to ~15 Ma. The horizontal buoyancy theory, modeled by England & Houseman (1986) and described by Dewey *et al.* (1989), would predict that the bulk of the Tibetan plateau stayed at a constant elevation above sea level or rose slowly from an elevation of 3 km or more during this period. Deformation was apparently concentrated in southern Tibet and the Himalayas (GTS, MCT, STD, and RZT) and transferred to the north (Tien Shari and Altai) and east (Red River shear zone). The STD extension in the High Himalaya, perhaps triggered by melting in the lower crust that also produced the Miocene leucogranites of the zone, probably lowered somewhat the elevation of southernmost Tibet (Burchfiel *et al.*, 1992). A Pliocene---Holocene normal fault system associated with underthrusting that is analogous to the STD system has been discovered in the Kongur Shari of the eastern Pamir (Brunei *et al.*, 1994),

The deformation of Tibet during this phase appears to be inhomogeneous, with N–S shortening in the southernmost part and some extrusion to the east. There is little evidence for N–S thrusting during this period, so one may conclude that the crustal thickness remained approximately constant. The youngest strata that are clearly folded by N–S shortening are interpreted as Miocene in age (Burg *et a/., 1983;* Dewey *et a/., 1988;* Burchfiel *et al.*, 1989). A planation surface developed in east-central Tibet, interpreted as formed during the Miocene, shows only minor warping and faulting that suggests little post-Miocene deformation (Shackleton & Chang, 1988). Dewey et *a/.* (1989) used an assumption about the initial angle between the conjugate strike-slip faults in Tibet and their present orientations with an assumption of horizontal plane strain (no net thickening) to estimate a minimum of -250 krn and maximum of --1160 km for the E–W extension and extrusion, with 29% and 40%, respectively, of N–S shortening of Tibet during the phase described in this section, If the extent of Tibet to the east of the eastern syntaxis (~95°E) is due to extrusion, then the --800 km from the syntaxis to the Longmen Shari (Figures 1 and 5b) would match an equivalent amount of extrusion,

2.5 **Late Tertiary--Quaternary extension**

A reconnaissance field expedition to Ulugh Muztagh in northernmost Tibet, which also crossed the margin of Tibet to the north (Molnar *et a/.,* 1987; Burchfiel *et a/.,* 1989; McKenna & Walker, 1990), concluded that crustal thickening of northern Tibet (at least near Ulugh Muztagh) happened before 10.5 Ma, according to geochemical interpretation of granitic intrusions as derived from partial melting of thickened continental crust, The younger volcanic rocks of Ulugh Muztagh are little deformed, so only a small amount shortening happened after 4 Ma, Active and Quaternary shortening is concentrated in the Altyn Tagh at the edge of Tibet, north of Ulugh Muztagh (Molnar *et al.*, 1987; Burchfiel *et al.*, 1989; McKenna & Walker, 1990).

The present extensional tectonics in Tibet results from the high gravitational potential causing horizontal buoyancy forces greater than the confining forces on its margins (e. g., Molnar & Tapponnier, 1978; England & Houseman, 1989; Molnar et *al., 1993).* Thickening of the crust and thinning of the mantle lithosphere act to both increase the gravitational potential and uplift the surface relative to sea level, therefore the change in the stress state within Tibet from compression ($\sigma_{xx} > \sigma_{zz}$) to extension ($\sigma_{zz} > \sigma_{xx}$) must have happened when the elevation reached a threshold, estimated from the distribution of focal mechanisms, of about 4500 m above sea level (Molnar *et al.*, 1993). This means that evidence of extension or major lithospheric thinning indicates that the plateau reached at least an elevation roughly 500 m below its present elevation.

The timing of the initiation of E–W extension in Tibet is under active study, primarily in a few areas of southern Tibet near Lhasa and in the Himalayas. Armijo *et al.* (1986) suggested that the active N–S trending normal faults formed after the late Pliocene at ~2 Ma, based on stratigraphic relations in the Yangbajian graben (-100 km NW of Lhasa). Isotopic cooling ages from the inactive, low-angle extensional detachment on the SE edge of the Nyainqentanglha range, which is the NW margin of the Yangbajian graben, record an earlier episode of extension in this area along a ductile shear zone (Pan & Kidd, 1992). Pan & Kidd (1992) interpret the timing of the ductile strain in the shear zone as occurring later than 11 Ma and before 5 Ma, and consider the Nyainqentanglha to be a metamorphic core complex similar to those found in the western US. The core complexes in the US represent the earliest stage of the Basin and Range extension, so this shear zone may represent the earliest extension in this part of southern Tibet (Pan & Kidd, 1992; Harrison *et a/.,* 1992a). Coulon *et a/. (1986)* suggested that potassic talc-alkaline Iavas of the Maquiang area in southern Tibet, dated at 15–10 Ma, indicate that extension might have initiated then. Coleman & Hodges (1995) determined a clear early age of -14 Ma for E–W extension in north-central Nepal, where the north-striking normal faults of southern Tibet cut through the Himalaya. Thus, it seems that E–W extensional stresses developed in southern Tibet before the middle Miocene (> 14 Ma; Coleman& Hodges, 1995).

The source of widespread, small-volume eruptions of basic and lesser felsic volcanic rocks in northern Tibet has been interpreted as coming from melting of lithospheric mantle (Arnaud *et a/.,* 1992; Turner et *al,* 1993). The geochemistry of the high-potassium basalts indicates they developed by partial melting of enriched subcontinental lithosphere. Two mechanisms for bringing necessary heat into the lithospheric mantle are removal of its lower part (England & Houseman, 1989; McKenna & Walker, 1990; Turner et *al.*, 1993; Molnar *et al.*, 1993) and intracontinental subduction (Arnaud *et a/.,* 1992). Lithospheric thinning is sometimes called delamination, but that term implies removal of an entire layer which is different from the partial removal of the mantle lithosphere modeled by England & Houseman (1989). Arnaud *et a/,* (1992) concluded that subduction of Asian lithosphere beneath the Kunlun was the most likely explanation of the basic and felsic volcanism in the westernmost part of northern Tibet, but agreed with the interpretation of the basic magmas in central northern Tibet as derived from lithospheric thinning or partial delamination (McKenna & Walker, 1990; Turner *et a/., 1993).* The ages of this basic volcanism extend back to 13 Ma, showing that lithospheric thinning should have started before this time in northern Tibet (Turner *et a/.,* 1993). Thus, both volcanism in northern Tibet and extension in southern Tibet are compatible with the uplift of the surface of Tibet to a high elevation, probably close to the present elevation, before 13–14 Ma. As noted by

Coleman & Hodges (1995) and Searle (1995), this is substantially earlier than the paleoclimatic indications of the initiation of the Asian monsoon.

The total amount of extension during this phase of the deformation history of Tibet is poorly constrained. Armijo *et al. (1986)* inferred the extension rate of one of the Quaternary graben systems (the Yadong-Gulu rift) based on field measurements, They then extrapolated that rate to the six other major graben systems in southern Tibet to produce a figure of 1±0.5 cm/yr for the total rate of extension over the 1100 km width cut by those fault systems. This amounts to about 2% E–W extension since 2 Ma, the time that they estimated for the initiation of extension. If this rate of -1 %/Ma extension were applicable to the entire time since 14 Ma, the total extension E–W would be nearly 14% or 140 km across the central portion of southern Tibet for this phase of deformation, Unfortunately, the uncertainties in this extrapolation are very large, probably more than a factor of two.

The graben formed by the mapped extension in southern Tibet, with the exception of the unique Yangbajian graben-Nyainqentanglha Shari (part of the Yadong-Gulu rift), have a relatively small amount of topographic relief, less than 500 m, and are quite widely spaced, so that the net extension of the south Tibetan crust is probably small, perhaps a total of 5%. For the rest of the plateau, the north and central parts, little deformation is apparent during Late Cenozoic, The degree of dissection is very low and surface manifestation of faulting is absent, as shown by the predominance of slopes less than 5°. The only areas of topographically expressed faults in northern Tibet are along the active Altyn Tagh and Kun Lun fault systems at the northern edge of the plateau. The peaks in the topographic relief at the north edge of Tibet in Figure 5A are related to these two fault systems.

3. *Erosion*

To compare the erosion history of Tibet with the uplift history, it is convenient to use the same divisions described above (§2). I begin with the modern erosional regime, because that is the period for which we have the most information. The total amount of erosion in an area is the integral of the erosion rate over a given period. The rate of erosion and dissection of an area is a function of (1) the relief that provides potential energy, (2) the amount and type of precipitation that does the geomorphic work, and (3) the erodibility of the materials exposed at the surface, The configuration of the drainage network strongly controls the distribution of erosion. Areas of internal drainage, such as the vast interior of Tibet, cannot lose mass through fluvial or glacial erosion so the net denudation will be approximately zero (except for eolian erosion), and medium-wavelength relief between highs and lows generally will be reduced by the redistribution of material.

The geologic evidence of denudation history for a given area generally falls into three categories: unconformities and truncation of deformed structures in the area, sedimentary deposits containing clasts eroded from the area, and thermochronometric measurements of cooling of rocks. The former two techniques are traditional geologic documentation of erosion, but do not directly provide measurements of erosion amounts and precise time control is sometimes difficult to obtain. The latter is a relatively recent geologic technique that can provide, in some cases, a detailed temperature vs. time history of a rock sample or samples by measuring the time since various minerals (or mineral domains) cooled enough to retain isotopes or fission tracks (e.g., Harrison et *al.*, 1992a; Copeland *et al.,* 1995). Mineral cooling ages can then be combined with assumptions (or measurements) of the geothermal gradient at those dates to estimate at what the depth below the surface were the samples at those times.

Erosional or tectonic denudation of the overlying rocks decreases the depth below the surface and causes the sample to cool and retain markers. Unfortunately, high rates of denudation, high local relief, or underthrusting of colder rocks can strongly affect or even temporarily reverse the normal geothermal gradient which greatly complicates the interpretation (e.g., Stüwe *et al.,* 1994; Copeland *et al.,* 1995).

3.1 Recent

The present erosion of Tibet is concentrated at the edges of the plateau, where there is high topographic relief (2000–6000 m; see Figure 5), steep slopes (Figure 4), and external drainage. Intense fluvial and glacial dissection maintains high local relief, The steepest slopes are in the Karakoram and Himalayas where deep valley slopes average about 35° and reach up to 78°. The along-strike averaged elevation of the High Himalaya is only slightly greater than that of the plateau, resulting from the combination of very high peaks and deep river valleys (Figure 5a; e.g., Wager, 1937; Bird, 1978). Dissection is more limited on most of the northern margin of the plateau in the Kunlun Shari where the topography drops 3-4 km over some 50–100 km into the Tarim and Qaidam Basins, but there are few deep valleys. In those regions, there is only a narrow band of steep slopes (Figure 4) and relief within swath profile (Figure 5a) reaches 3 km compared to the 5.5 km relief in the Himalayas. The areas of high relief (and slope) clearly correlate with the areas of higher precipitation (Figure 8).

The Karakoram is deeply dissected by intense Quaternary and continuing modern glaciation, producing deep valleys between high peaks, with an average range elevation slightly higher than that of Tibet (Figure 5b). The elevation of the peaks in the Pamirs is about the same as in the central part of Tibet, but the average elevation (3.9 km) is lower by about 900 m due to a much higher degree of dissection, removal of mass by erosion, or thinner crust (Figure 5b). The western Pamirs are much more dissected than the eastern portion that includes a relatively flat area at about 4.2 km elevation (Figures 3 and 4). The increase in erosional relief towards the west in the Pamirs matches an increase in annual precipitation (Figures 5b, 8).

As described above (§2), the eastern part (nearly a third) of the Tibetan plateau slopes gradually downward from the center of the plateau beginning roughly at the external drainage divide (Figure 5b). Both the mean and maximum elevations decrease, but the minimum elevations drop faster due to the deep canyons that eventually drain into many of the major rivers of south and east Asia, including the Brahmaputra, Mekong, Yangtze, and Huang Ho (Yellow) Rivers, This produces the gradual eastward increase in relief shown in Figure 5b. The more gradual long-wave length rise of eastern Tibet also allows precipitation to penetrate farther onto the plateau (Figure 8). The greater precipitation and external drainage are then responsible for the greater dissection and relief. Since eastern Tibet is likely to have been formed by extrusion, the external drainage of this area is probably an old feature, perhaps dating back to the Miocene or earlier.

Southern Tibet includes a substantial area of external drainage, where the Indus, Ganges, and Zangbo/Brahmaputra and tributary rivers leave the plateau (Fig, 1). This zone is also somewhat wetter (mostly 400-600 mm/yr) than the central and northern parts of Tibet, but much drier than the Himalaya just to the south (Fig. 8). The Zangbo River that drains southern Tibet is now at elevations of 3.5–4.5 km above sea level, so the regional base level is very high. Sand grains from a dune next to the Zangbo (Tsangpo) River were analyzed by Cope[and & Harrison (1990). They found that the K-feldspar grains in the sand had age minima with a maximum at -17 Ma and no grains younger than 8 Ma, and suggested that all of these grains could have come from the erosion of the

Gangdese batholith, which is north of the river. While this sample of 26 grains may not be representative of the large region of southern Tibet and the north slope of the Himalaya that drains through this point, the lack of very young ages suggests there is little or no area of rapid, deep erosion in that region, unlike the southern slope of the Himalaya (Copeland & Harrison, 1990),

The climate of the central part of Tibet is now arid (less than 400 mm of precipitation per year; Figure 8) and cold (average annual temperature below freezing; Climatic Atlas of Asia, 1981), so the rate of fluvial erosion is presumably low, The highest peaks (above about 6500 m) are presently glaciated. However, due to the aridity of Tibet, Pleistocene glaciers did not extend much below this altitude (Wissmann, 1959; Derbyshire *et a/.,* 1991), In this cold, dry desert regime, erosional processes presumably are less efficient at transporting material over long distances and thus slowly remove topography uplifted by tectonic deformation. The disconnected internal drainage basins of Tibet also prevent long-distance transport of material (except for eolian erosion). In contrast, the Karakoram, Himalaya and eastern Tibet receive a much greater amount of precipitation than central Tibet, 600--4000 mm/yr over much of the area (Figures 5 and 8); the spectacular and extensive glaciers of the Karakoram are maintained by this moisture. The clear spatial correlation between the areas of deep dissection and present high precipitation (Figures 4, 5, and 8) indicates that this climatic distribution has continued for a substantial period, at least throughout the Quaternary, and perhaps the late Tertiary.

3.2 **Mesozoic-Early Tertiary**

The widespread deposition of marine limestones over much of Tibet during the Albian/Aptian indicates that those areas were below sea level (§2.2) and presumably not being eroded significantly at that time, The volcanics of the Gangdese-Karakoram arc were at least locally eroded during this time to provide the volcanic clasts in the Takena formation (also Albian/Aptian) on the northern part of the Lhasa block (Coulon *et a/.,* 1986). Erosion of the active arc was also providing the primary source for the sedimentary rocks in the forearc basin to the south (Xigaze series) during the Albian/Aptian. During the late Cretaceus, at least the northern parl of the Lhasa block was more strongly eroded as shown by the truncation of structures in the Takena formation and angular unconformity with the overlying upper Cretaceous--Paleocene Linzizong strata (*e.g.,* Burg *et al,* 1983; §2.2). The Linzizong volcanics have at least one old apatite fission-track age near Lhasa (39±6 Ma) that indicates that sample has not been buried more than ~3 km since the Eocene (Pan *et al.*, 1993). The thick deposits and apparent lack of deep burial of Linzizong volcanics suggest little erosion of the northern Lhasa block since their deposition. Not much else is known about erosion in the rest of Tibet during this period, but the widespread outcrop of Triassic to middle Cretaceus sedimentary rocks in Tibet precludes large amounts of denudation (Kidd *et a/.,* 1988; Dewey *et a/.,* 1988). Dewey *et a/.* (1988) asserted that net denudation of Tibet north of the Gangdese arc has been -2 km, which may be too small to detect even with low-temperature thermochronometry such as apatite fission tracks.

3.3 **Early Tertiary**

Sometime in the Eocene, presumably after the initial contact between the Indian continent and Tibet, coarse conglomerates were deposited on the southern margin of the Gangdese arc that include many clasts derived from the erosion of the arc (e.g., Heim & Gansser, 1939; Burg *et a/.,* 1983). The large thickness of some of these molasse deposits suggests significant erosion of the volcanic arc, The Late Cretaceous--early Tertiary sedimentary and volcanic rocks in southern Tibet appear not to have been buried to depths greater than --6 km (Coulon *et a/.,* 1986; Copeland et *a/.,* 1995), so at least those parts were neither the sites of deep deposition nor denudation. Coarse redbeds of Eocene age were also deposited in at least a few sites in east-central Tibet (Kidd *et a/.,* 1988; Leeder *et a/.,* 1988), which suggests an adjacent area of erosion. Other elastic deposits are of unknown age (Kidd *et a/.,* 1988). Volume balancing of the sediment volumes in the Bengal and Indus fans and the foreland basin of the Himalaya compared to estimates of eroded volume of the Himalaya has been used to infer a significant sediment contribution from Tibet before 25 Ma (Johnson, 1994), but this sediment was probably derived primarily from the Gangdese arc.

Some of the plutonic rocks of the Gangdese arc show post-crystallization cooling, presumably related to erosion of the overlying rocks, between -50 Ma and -30 Ma (Copeland *et a/.,* 1987; 1995). For example, $^{40}Ar/^{39}Ar$ analyses of hornblende, biotite, and K-feldspar from samples of the Dagze granite, just south of Lhasa, indicates very rapid cooling from >500°C to ~275°C in -1 my. at about 55 Ma that is most likely to be the time of pluton crystallization (Copeland *et a/,,* 1995). Detailed modeling of K-feldspar from one sample shows the Dagze granite stayed at a constant temperature from -55 to -44 Ma, then cooled relatively rapidly from -44 Ma to 40 Ma at a rate of 20 °C/m.y. to reach 215°C (Copeland *et* a/., 1995). Assuming a moderately high (and constant) geothermal gradient of 30 °C/km for that time, one can estimate a denudation rate of ~0.66 mm/yr and denudation amount of 2.6 km for that granite. Other Gangdese plutons show slower rates of cooling during the Eocene, and Copeland *et a/.* (1 995) reported denudation rates for plutonic rocks since 40 Ma between 0.18 and 0,30 mm/yr. Since there is a lack of evidence for tectonic denudation during this time, these moderate denudation rates are probably equal to erosion rates in the Gangdese belt of southern Tibet,

3.4 **Middle Tertiary**

An episode of rapid cooling, and presumably rapid erosion, affected many of the Gangdese plutons of southern Tibet in the Miocene (Harrison *et a/.*, 1992a; Pan *et a/.,* 1993; Copeland *et a/.,* 1995). Across east-central Tibet along the Geotraverse route, Shackleton & Chang (1988) describe evidence for extensive mid-late Miocene planation, although their timing control is less precise. The area studied in the most detail so far is the -225-km-long segment of the Gangdese arc near Lhasa. At almost every site, there is a distinct difference between the cooling due to emplacement and that due to unroofing, so the rapid cooling in the Miocene is interpreted as being due to denudation. Unfortunately, detailed thermal and stratigraphic chronologies have not yet been obtained for the enormous interior of Tibet, so little can be determined about erosion there.

Pan *et a/,* (1993) and Copeland *et a/.* (1995) report thermochronometric data from six plutons that clearly show rapid cooling after 40 Ma. Four of these experienced a brief period of extremely rapid cooling sometime in the interval 26-15 Ma (Copeland *et al., 1995).* One well-constrained swift rate is on the northern part of the Quxu pluton that cooled at >80°C/m.y. at -1 7–15 Ma according to the $^{40}Ar/^{39}Ar$ age of biotite and the fission track age of apatite (Pan *et a/.,* 1993). This cooling is much younger than the intrusion age of the Quxu pluton that is --42-50 Ma (Copeland *et a/.,* 1995). Multi-domain modeling of K-feldspar samples from different elevations at Quxu allows the calculation of the paleogeothermal gradient of -38 °C/km during the interval 26–20 Ma (Copeland *et a/.,* 1995). If this high gradient can be applied to the 17-15 Ma period at Quxu and the geothermal gradient remained constant, the cooling rate would register ~2 mm/yr and a total of -4 km of denudation during that time. Lower geothermal gradients would require higher rates of erosion to produce the same cooling rate. The Pachu pluton some 100 km to the west of Quxu also cooled at a -80 °C/m.y. rate between 14.1 and 11.7 Ma according to $^{40}Ar/^{39}Ar$ ages on biotite and K-feldspar modeling, but no clear intrusion age has yet been determined for it (Copeland *et* al, 1995). If the intrusion is assumed to be >40 Ma, then the rapid cooling is most likely due to denudation, as for the Quxu pluton.

Rapid denudation rates can be generated by either erosional or tectonic unroofing, The rapid exhumation of the Gangdese batholiths during the Miocene is likely to be related to the motion on the Gangdese Thrust System (Harrison *et al,* 1992a; Yin *et a/.,* 1994). Since the GTS was initiated at 27±1 Ma (Yin ***et al.,*** 1994), it appears that the fastest denudation rates lagged behind the movement on the thrust system. The two samples that show earlier rapid cooling -26 Ma are both near the southern margin of the Gangdese and the GTS, and may have been initially cooled by colder rock being underthrust beneath them (Yin *et* ***a/.,*** 1994). The general trend of younger ages of the rapid cooling pulse northwards could be explained ‘by the southward movement of the Gangdese hanging wall over a footwall ramp in the GTS (Copeland eta/., 1995). This trend could also be caused by a northward-propagating erosional front. The extensional denudation due to movement on the STD, at roughly the same time, occurred some 150-200 km to the south, so it is unlikely to have affected the Gangdese batholiths and the rapid Miocene denudation was probably caused by fast erosion of the overlying rocks (Copeland eta/., 1995).

The rapid erosion of the Gangdese batholiths during the Miocene has been interpreted as evidence that the surface of southern Tibet was uplifted rapidly relative to sea level during the late Oligocene–early Miocene by isostatic compensation of major crustal thickening on the GTS (Harrison eta/., 1992a; Pan eta/., 1993; Yin *et a/., 1994;* Copeland *et al.*, 1995). Another possibility (argued against by Harrison and coworkers) is that some or most of the >4 km of Miocene erosion recorded by the cooling of rocks is due to deep erosion down through an already elevated terrain, which would mean uplift of rocks relative to sea level but a lowering of the average surface elevation relative to sea level. There is evidence in the $\delta^{18}O$ record from deep-sea sediments in the Atlantic that the global climate cooled significantly in a short period near 15 Ma (Raymo & Ruddiman, 1992). This cooling might have contributed to increased erosion rates in southern Tibet without any change in the surface elevation.

In addition, the Miocene extensional movement on the STD is likely to have modified, and probably increased greatly, the gradient of the rivers draining southward from the Gangdese across the Himalaya and STD by lowering the regional base level. This would require that the Gangdese be already at a high elevation, [f the STD also acted to lower the average elevation of the Himalaya, it would have allowed more precipitation to reach the Gangdese by reducing the rain shadow effect, Increased river gradients and more precipitation would empower more rapid erosion, The Gangdese is now at an average elevation 5–5.5 km above sea level (Figure 5a), but it is not rapidly eroding due to low amounts of precipitation and high regional base level that are both caused by the Himalaya to the south ($3.1; Copeland & Harrison, 1990). Thus, the rapid erosion of the Gangdese during and after the Miocene may have been controlled more by the complex history of uplift in the Himalaya than by regional uplift of southern Tibet in the Miocene.

3.5 Late Tertiary–Quaternary

The present morphology reflects the Late Cenozoic erosional history of Tibet (§1 .2) to a large extent. For most of the plateau, especially the north and central parts, erosion has apparently been very slow for much of the Late Cenozoic. The degree of dissection is very low, as shown by the predominance of slopes less than 5° (Figure 4). Many of the scarps of the graben systems in Tibet are only slightly eroded (Masek *et al.*, 1994b). The drainage network is very poorly developed in the dry climate of interior Tibet (Figure 8). In contrast, the deep canyons and steep slopes in the externally drained eastern Tibet, Himalaya, and Karakoram probably represent rapid erosion during all the Quaternary and perhaps some of the Pliocene.

In a few sites of extremely rapid erosion, thermochronometry can provide detailed information about the past few Ma. For most areas in Tibet, however, the rocks exposed at the surface today probably were already too shallow to record the erosion of the overlying rocks. If one assumes an apatite closure temperature of -100”C and a high geothermal gradient of 40 °C/km, then more than 2.5 km of rock must be removed before apatite fission tracks will show resetting. For example, at locations where the erosion rate is <0.5 mm/yr, there will be little or no information about changes in the erosion rate since 5 Ma, but one can still calculate an average denudation rate for the time since the rocks cooled through the apatite closure temperature (assuming a geothermal gradient).

A comprehensive review of thermochronologies in the Himalaya and Karakoram is beyond the scope of this paper, but a few examples will serve to illustrate the evidence for rapid denudation in those areas. One site where there is clear evidence of extremely swift Late Tertiary-Quaternary denudation is the Nanga Parbat-Haramosh Massif in the western Himalayas where rates of 1–8 mm/yr have been calculated for various intervals since 4 Ma from fission track and other isotopic ages (e. g., Zeitler, 1985; George *et a/.,* 1995), Exhumation rates as high as this certainly require caution to interpret (Stüwe *et a/.,* 1994), but rates >2 mm/yr are required by the data, which include an apatite age as young as 0.4 Ma where the Indus River gorge cuts the massif (Zeitler, 1985). This area has very steep slopes (averaging >30 degrees; Figure 4) that are both a consequence and a cause of the rapid denudation, and it has reasonably high annual precipitation (>1 200 mm/yr; Figure 8). The extreme rates in this area are probably related to the structural syntaxis at Nanga Parbat, combined with the abundant erosional power of the Indus River.

High denudation rates have also been determined in the Karakoram. At K2, the highest peak in the Karakoram, samples collected over 3300 m of elevation range have apatite fission track ages between 2 Ma and 3 Ma (Foster *et al.*, 1994). Horizontal temperature gradients are likely in this region of extreme relief (Figure 5b) and a one-dimensional cooling assumption is probably invalid (Stüwe *et al.*, 1994), but the concordance of ages over that vertical range requires fast denudation (>1 mm/yr) during the Pliocene and Quarternary. A minimum of 34 km and possibly as much as 6 km of rock must have been removed from the average surface elevation, now at ~5.5 km (Figure 5b; Foster *et al.,* 1994). Even more rapid cooling (-60 °C/m.y.) is required by an $^{40}Ar/^{39}Ar$ K-feldspar model age of 4.3±0.4 and closure temperature of ~260°C for a sample of a leucogranite in the Karakoram some 60 km west of K2 (Schärer ***et al.,*** 1990). Most or all of this exhumation in the Karakoram is likely due to swift erosion.

Uncertainties about geothermal gradients can be largely avoided by using thermobarometry and geochronology together. For example, peak metamorphism of rocks now exposed at the surface in the Zanskar High Himalaya occurred at depths of 25–27 km at a time inferred >25 Ma (Searle ***et al.,*** 1992). This translates to average exhumation -1 mm/yr since the Miocene, although some component is likely due to tectonic denudation by the Zanskar shear zone, and not pure erosional unroofing. Apatite fission-track ages from the Kishtwar Window in the Zanskar Himalaya are as young as 1 Ma, which requires very rapid Quaternary unroofing, but most of Zanskar has apatite ages of 6–7 Ma (Sorkhabi, 1993). Other results from the Higher Himalaya reviewed by Sorkhabi & Stump (1993) indicate denudation amounts of at least 10 km since 17 Ma, or minimum average exhumation rate of 0.5 mm/yr. Other areas on the margins of

Tibet, in the Himalaya, Karakoram, eastern Tibet, and perhaps the Kunlun, that have steep short-wavelength slopes, external drainage, and relatively high precipitation arelikely tohave similar high (>0.5–1 mm/yr) rates of denudation during at least the last few Ma, but there is also likely to be substantial variation in rates over both time and space.

Some of the detailed thermochronologies of southern Tibet show cooling ages from the late Miocene--Pliocene (Pan *et al.,* 1993; Copeland *et al.,* 1995). All the young cooling ages in southern Tibet, from rocks north of the Himalayan leucogranites, are interpreted as denudation ages not intrusion ages. Some of the young denudation is clearly related to extension on the faults of graben systems, such as the Nyainqentanglha shear zone (§2.4; Pan & Kidd, 1992). Samples from the footwall of the Nyainqentanglha shear zone yield apatite fission track ages as young as 3.3±0.6 Ma (Pan *et al.,* 1993), which converts to an average cooling rate of 30±10°C/m.y. since that time assuming an apatite closure temperature of 100±20°C. Combining the apatite ages with K-feldspar $^{40}Ar/^{39}Ar$ ages from the same samples produces cooling rates of 30–300°C/m.y. for intervals between 9.5 anti 3.3 Ma (Pan *et al.,* 1993). Even with a very high geothermal gradient, these late Miocene–Quaternary cooling rates convert to denudation rates of 0.5–5 mm/yr, which are most likely due to tectonic exhumation of a core complex in this part of the Yadong-Gulu graben system (Pan & Kidd, 1992). One other thermochronology in southern Tibet shows an apatite fission track age of 6.9±0.9 Ma, but the majority of apatite ages are older than 15 Ma (Pan *et al.,* 1993; Copeland et *al.*, 1995). This is consistent with the K-feldspar $^{40}Ar/^{39}Ar$ ages in the modern Zangbo sand that peak at -17 Ma and have a minimum of 8 Ma (Copeland & Harrison, 1990). This evidence suggests that there has been less than about 3 km of denudation and erosion in nearly all of southern Tibet since ~15 Ma, except where tectonic unroofing by extension has occurred.

Detailed thermochronologies are not yet available from the interior of Tibet, and because the expected erosion rates there are low, measurements would be unlikely to provide much information on the Pliocene–Quaternary denudation, One location at the northern edge of Tibet in the Kunlun close to Golmud was studied by Lewis (1990), who measured apatite fission-track ages of -20 Ma and track-length evidence that the rocks were in the partial annealing zone (50–100°C) before 20 Ma. It is possible that as little as 2 km has been eroded there in the past 20 Ma for an average rate of 0.1 mm/yr. There is a good chance that the rest of the internally drained, dry interior of Tibet has even slower Late Cenozoic erosion than the Kunlun. An upper-bound average erosion rate of 0.1 mm/yr would remove only 1.5 km from the mean surface of Tibet in the past 15 Ma, which is roughly the same amount of crustal thinning as 2% extension of the 70 km thick crust.

4. ***Summary: Tibet history***

The evidence available now and discussed above (§2 and §3) is still insufficient to rigorously constrain the uplift and erosion histories of the Tibetan Plateau, but it is perhaps useful to summarize by synthesizing the data into a speculative combined history. At each of the major stages in the history, an isostatic calculation of the expected elevation (see Appendix) for the inferred crust and upper mantle structure is shown in Figure 7. The times of the stage boundaries plotted (100, 45, 25, and 14 Ma) are not exact, but provide rough times for which the lithospheric columns and elevations may apply. There are surely variations in the lithospheric structure in Tibet today (e.g., Molnar, 1988) and in the past, but the uncertainties in reconstructing the past are large enough that only on'e "Tibet interior" column is shown for each stage, The details of this history (elevations and times) are unlikely to be correct, but the overall features of the elevation history illustrate the implications of the deformation of the plateau.

The early history of Tibet, before the impingement of the Indian continent, is difficult to constrain because later events have overprinted the earlier structure. One paleoelevation is reasonably well determined by the Albian-Aptian marine limestones that indicate that much of Tibet was slightly below sea level at ~100 Ma (see §2.2). I assume a slightly thin continental crust of 30 km thickness with a medium-thickness mantle lithosphere of 100 km to produce an isostatic elevation of -0.2 km below sea level (Fig. 7). This seems reasonable for the lithospheric structure of recently accreted terranes, but other structures fit the same isostatic elevation constraint.

Moving forward in time through the mid-late Cretaceus and Paleocene history, it seems likely that there was some shortening of the Tibetan crust, perhaps due to an episode of flat- or shallow-slab subduction. Distributed shortening of ~23% (thickening factor 1.3) is assumed to be vertical plane strain in both the crust and mantle lithosphere to produce a crust 39 km thick and mantle lid 130 km thick that has an isostatic elevation of 0.3 km above sea level just before the contact of the Indian continent (Fig. 7). This elevation is consistent with the lack of marine deposition in Tibet at -45 Ma since it is significantly above sea level, but there are no constraints on the elevation at this time ($2.3). If one assumes that there was no shortening of Tibet before the collision of the Indian continent, then this lithosphere configuration can still apply to later time in the shortening history after the collision.

Distributed shortening and thickening of the crust and mantle lithosphere of Tibet is assumed to continue from the collision to the Miocene to arrive at a crustal thickness of 73.5 km for the pre-extension-phase Tibetan Plateau (see below). Starting at a 39 km crustal thickness requires 47% shortening (thickening factor 1,88), while the total thickening from a 30 km crustal thickness is 59% (a factor of 2.45), The mantle lithosphere after the same amount of shortening would have a thickness of 245 km before any lithospheric thinning. The isostatic surface elevation for this lithospheric structure would be -2.2 km above sea level in the early Miocene, -25 Ma (Fig. 7). This elevation depends on the amount of shortening of the lithosphere, the initial configuration (especially the ratio of the thickness of the crust to mantle lid), and the assumed densities.

The maximum elevation and consequent change to an extensional regime of the Tibetan Plateau are likely to have resulted from some type of lithospheric thinning process such as convective removal of the thermal boundary layer (e.g., England & Houseman, 1989). Extension and volcanism, which were presumably related to this lithospheric thinning, indicate that it happened before 14 Ma (§2.5). The apparently sudden development of the Gangdese Thrust System, Main Central Thrust, and South Tibetan Detachment in southernmost Tibet and the Himalaya at -25 Ma record a change in the deformation regime of Tibet that suggests that the lithospheric thinning may have begun then, I speculate that the crust of the interior of Tibet between 25 Ma and 14 Ma remained about the same thickness, since it appears that deformation was concentrated in southern Tibet and the Himalaya between -25 and -14 Ma, while the mantle lithosphere was thinned dramatically from ~245 km to ~105 km thickness sometime during the Miocene. This amount of thinning would result in about 3.2 km of uplift of the surface of Tibet (Fig. 7), similar to the calculations of England & Houseman (1989).

As described in $2.1, the one exact constraint on the elevation of Tibet is the present 5.02 km above sea level. I assume a crustal thickness of 70 km and mantle lithosphere thickness of 100 km for a point in central Tibet (see Appendix for references). To estimate the Late Cenozoic elevation history, we can work backwards from the present. If there has been roughly 5% extension of Tibetan crust since -14 Ma ($2.5), then the crust would have been -73.5 km thick before extension. As discussed in §3.5, the amount of erosion in central Tibet during this time is likely to be small and less than the uncertainty in the amount of extension, so I neglect any erosional thinning of the crust. If the crust has been extended, then presumably the mantle lid below was also extended by roughly the same 5% and would have been 105 km thick at ~14 Ma. This lithospheric structure would have an isostatic elevation of -5.4 km above sea level (Fig. 7) The -400 m higher than present elevation represents a likely upper bound to the highest elevation that the Tibetan Plateau reached before its latest extensional phase, probably in the Miocene.

How does the present low-relief morphology of Tibet match with this proposed history? At moderate-to-long wavelengths from 10 km to hundreds of km, central Tibet is extremely flat (Figures 4 and 5; Fielding *et al.*, 1994). This lack of long-wavelength relief may be caused by flow in the lower crust due to the extreme thickness of and heat production within the Tibetan crust (Gaudemer *et a/.,* 1988; Bird, 1991) or other processes that result in a low effective viscosity of the lithosphere (England & Houseman, 1986). Flexural modeling of the rift flank uplift on some of the graben in Tibet indicates a very thin effective elastic plate thickness (<1 O km) that would be consistent with a low viscosity zone in the Tibetan crust (Masek *et* a/., 1994b). The low rates of fluvial cut-and-fill processes for central Tibet (described above in $3.1) cannot erase tectonic relief at moderate wavelengths, so either there has been little significant deformation of the upper crust in the late Cenozoic or some other process, such as viscous flow in the lower crust, has removed relief at these wavelengths.

Another major factor is time of erosion since the last construction of tectonic relief, The short-wavelength topographic slopes of central Tibet are generally low but locally high close to the young normal faults of southern Tibet (Armijo et ***a/.,*** 1986). The relatively small scale of these grabens and their scattered distribution suggests that they are a minor upper crustal deformation and do not indicate significant lithospheric deformation. Fault scarps formed by youthful deformation, if they were present in the large smooth areas of northern Tibet, would probably be well preserved and visible at the surface as they are in other regions of similar climate such as the thrust belt on the southern margin of the Tien Shari, southern Tibet or the Basin and Range of the Western USA. The late Cenozoic deformation regime of Tibet cannot involve signification deformation, especially shortening, of the surface of northern Tibet. Early Cenozoic shortening of the upper crust of Tibet can be consistent with the present relief since there has been enough time for even slow erosion to have reduced the constructional topography.

Some models of denudation have made a simplifying assumption that erosion rates are directly proportional to elevation. This may be reasonable in a wet, narrow, externally drained mountain range such as the Southern Alps or Taiwan, but it clearly does not apply to wide plateaux such as Tibet or the Altiplano-Puns (Fig. 2; Isacks, 1992). The great interior of Tibet is now at (and probably has been for millions of years at) a very high elevation (-5 km), but has a low erosion rate. Since the distribution of precipitation strongly affects erosion rates, the intense focusing of precipitation at moderate elevations (-2 km) on a mountain front such as the Himalaya may cause the most rapid erosion at elevations below the peak of the range (Masek *et al.*, 1994a). In addition to the spatial variations of climate and drainage network that produce variations in erosion rates at a given time, there have been major temporal changes in global climate, especially during the Cenozoic, that surely affect erosion rates (Molnar & England, 1990). These considerations (and others) prevent a direct relationship between erosion and uplift of the surface with respect to sea level (England & Molnar, 1990).

5. *Conclusions*

The surface of Tibet records the integrated effects of tectonic and erosional processes over at least 20 million years, but it is most sensitive to more recent effects that have tended to erase evidence of earlier processes. There is still little age control on tectonics within central Tibet, although data has been collected recently in the more accessible areas of the plateau, especially near the Golmud–Lhasa road (see §2). At some of those sites, there is evidence of shortening within the upper crust at various times before ~15 Ma ($2.2–2,4), This suggests that some amount of distributed shortening affected the Tibetan lithosphere, probably starting in the Early Cenozoic after the coupling of the Indian continent with Tibet. The speculative model presented in Figure 7 shows a scenario where lithospheric thickening (vertical plane strain) by a factor of two or more may have resulted in a crust slightly thicker than the present crust of Tibet some time during the Miocene. Thickening of the mantle lithosphere by the same amount would weigh down the surface elevation to less than half of the present elevation (Fig, 7). If a large part of this overthickened mantle lithosphere was then rapidly removed, then the plateau would rise quickly to an elevation perhaps slightly higher than today (England & Houseman, 1989), Rapid Miocene erosion of the Gangdese arc was probably caused partly by deformation of southern Tibet (Harrison *et* a/., 1992a; Copeland *et a/.,* 1995), but it may have been facilitated in part by uplift of the plateau and in part by changes in base level and precipitation caused by Miocene events in the Himalaya. The youngest deformation within Tibet is clearly the E-W extension that has been interpreted as beginning as early as 14 Ma by Coleman & Hodges (1995; §2.5). This extension of the Tibetan lithosphere may have caused the elevation to decrease somewhat in the past several million years, unless crustal thinning was balanced by some other process, Some early studies suggested that the Tibetan plateau is continuing to shorten in a N–S direction by "accordion-style" thrust faulting and folding (e.g., Dewey & Burke, 1973). The present smooth topographic surface cuts across upper Tertiary folded strata without noticeable relief, despite apparently low erosion rates. Folding and thrust faulting of the interior of Tibet, therefore, has not been active during latest Tertiary and Quarternary time. Uplift of the plateau during the late Cenozoic must have been through a mechanism that does not involve significant shortening of the uppermost crust, such as the underthrusting of India beneath Tibet (e.g., Barazangi & Ni, 1982; Beghoul *et* al., 1993) or lithospheric thinning (e.g., England & Houseman, 1989). Distributed shortening may well describe the early evolution of the Tibetan plateau, before the change to an extensional regime at, perhaps, 14 Ma. The symmetrical and extremely narrow peak of the hypsometry over an area of more than half a million square km in north- and south-central Tibet (Figure 5) suggests some kind of fluid crustal process is operating recently to level the elevation (England & Houseman, 1986), perhaps the lower crustal flow of Bird (1991).

Appendix: Isostatic calculations

Isostatic calculations to balance the mass of crust and mantle lithosphere columns are similar to balancing structural cross-sections in that they can prove that particular interpretations are false but cannot prove that others are correct, only feasible. As described in $2.1, local (one-dimensional) isostasy is a reasonable approximation for the interior of Tibet because flexural support of mass anomalies will be small due to the enormous width of the plateau. The method used here is to choose a reference column with the present elevation, crust and mantle lithospheric structure of Tibet, calculate its total mass down to a depth of 300 km below sea level, and then match the mass of columns for other times to that reference column, The densities of the crust, mantle lid, and asthenosphere are even more difficult to measure than thicknesses, so I assume that the density contrasts between the crust and asthenosphere (425 kg/m^3) and between the mantle lid and asthenosphere (75 kg/m^3) are constant to simplify the isostatic calculations (Fig. 7).

A reasonably well accepted value for the present thickness of the crust in most of Tibet is -70 km (e.g., Molnar, 1988; Him, 1988), but the thickness of the mantle lithosphere (lid) is more controversial and variable in different parts of the plateau. Seismic measurements of P_n propagation for southern Tibet have produced estimates of a thick, cold mantle lid: >100 km (Holt & Wallace, 1990) and 135–1 80 km thick (Beghoul *et a/.,* 1993). Other seismic measurements indicate a relatively hot, thin mantle lithosphere in north-central Tibet, which would require a thinner crust to have the same elevation (e.g., Bourjot & Romanowicz, 1992; Molnar *et a/., 1993).* I assume a medium thickness of 100 km for the isostatic calculations done here, The masses of the other lithospheric structures are then matched against this assumed modern configuration to determine their isostatic elevation (Fig. 7).

Acknowledgment ts

The author would like to thank J.-P. Burg, D. Spencer, and C. Spencer-Cervato for organizing the outstanding workshop at Monte Verità, Ascona, Switzerland, and for inviting me to attend, Some of the ideas expressed in this paper were formed by the many discussions and lectures at that workshop. P. Molnar, P. England, and W. Kidd gave some helpful suggestions. Part of this research was performed at the Jet Propulsion Laboratory, California Institute of Technology, under contract with NASA,

REFERENCES

Argand, E., 1924, La tectonique de L' Asie: Comptes Rendus, Congrès Géologique International, v. 1, p. 171-372.

Armijo, R., Tapponnier, P., & Tonglin, H., 1989, Late Cenozoic right-lateral strike-slip faulting in Southern Tibet: Journal of Geophysical Research, v. 94, p. 2787–2838.

Armijo, R., Tapponnier, P., Mercier, J.L. & Tong-Lin, H., 1986, Quaternary extension in southern Tibet: Field observations and tectonic implications: Journal of Geophysical Research, v. 91, p. 13,803–1 3,872.

Arnaud, N. O., Vidal, P., Tapponnier, P., Matte, P., & Deng, W. M., 1992, The high K_20 volcanism of northwestern Tibet: Geochemistry and tectonic implications: Earth and Planetary Science Letters, v. 111, p. 351–367.

Barazangi, M., & Ni, J., 1982, Velocities and propagation characteristics of Pn and Sn beneath the Himalayan arc and Tibetan plateau: Possible evidence for underthrusting of Indian continental lithosphere beneath Tibet: Geology, v. 10, p. 179--185.

Beghoul, N., Barazangi, M., & Isacks, B. L., 1993, Lithospheric structure of Tibet and western North America: Mechanisms of uplift and a comparative study: Journal of Geophysical Research, v. 98, p. 1997–2016.

Bird, P., 1978, Initiation of intracontinental subduction in the Himalaya: Journal of Geophysical Research, v. 83, no. B1 O, p. 4975-4987.

Bird, P., 1991, Lateral extrusion of lower crust from under high topography, in the isostatic limit: Journal of Geophysical Research, v. 96, no. B6, p. 10,275–10,286.

Bourjot, L., & Romanowicz, B., 1992, Crust and upper mantle tomography in Tibet using surface waves: Geophys. Res. Lett., v. 19, p. 881-884.

Bowin, C,, 1991, The earth’s gravity-field and plate-tectonics: Tectonophysics, v. 187, p. 69–89.

Brias, A., Patriat, P., &Tapponnier, P., 1993, Updated interpretation of magnetic anomalies and seafloor spreading stages in the South China Sea, implications for the Tertiary tectonics of SE Asia: Journal of Geophysical Research, v. 98, p. 6299-6328.

Brunei, M., Arnaud, N., Tapponnier, P., Pan, Y., & Wang, Y., 1994, Kongur Shari normal fault: Type example of mountain building assisted by extension (Karakoram fault, eastern Pamir): Geology, v. 22, p. 707-710.

Burchfiel, B. C., & Royden, L. H., 1985, North-south extension within the convergent Himalayan region: Geology, v. 13, p, 679–682,

Burchfiel, B. C., & Royden, L, H., 1991, Tectonics of Asia 50 years after the death of Emile Argand: Eclogae geol. Helv., v. 84 (3), p. 599–629.

Burchfiel, B. C., Chen Z., Hodges, K. V., Liu Y., Royden, L. H., Deng C., & Xu J., 1992, The South Tibetan Detachment System, Himalayan orogen: Extension contemporaneous with and parallel to shortening in a collisional mountain belt: Geol. Sot, Amer. Special Paper no. 269, 41 p.

Burchfiel, B. C., Molnar, P., Zhao Z., Liang K., Wang S., Huang M., & Sutter, J., 1989, Geology of the Ulugh Muztagh area, northern Tibet: Earth and Planetary Science Letters, v. 94, p. 57–70.

Burchfiel, B. C., Zhang P., Wang Y., Zhang W., Song F., Deng Q., Molnar, P., & Royden, L., 1991, Geology of the Haiyuan fault zone, Ningxia-Hui Autonomous Region, China, and its relation to the evolution of the northeastern margin of the Tibetan plateau: Tectonics, v. 10, no, 6, p. 1091-1110,

Burg, J.-P., Davy, P., & Martinod, J., 1994, Shortening of analogue models of the continental lithosphere: New hypothesis for the formation of the Tibetan plateau: Tectonics, v. 13, p. 475-483.

Burg, J.-P., Proust, F., Tapponnier, P., & Chen G. M., 1983, Deformation phases and tectonic evolution of the Lhasa Block (southern Tibet, China): Eclogae geol. Helv., v. 76, p. 643–665.

Climatic Atlas of Asia, 1981, Maps of mean temperature and precipitation: Geneva, World Meteorology Organization, 28 p.

Coleman, M., & Hodges, K., 1995, Evidence for Tibetan plateau uplift before 14 Myr ago from a new minimum age for east–west extension: Nature, v. 374, p. 49–52.

Copeland, P., Harrison, T. M., Kidd, W, S, F., Ronghua, X., & Yuquan, Z., 1987, Rapid early Miocene acceleration of uplift in the Gangdese Belt, Xizang-southern Tibet, and its bearing on accommodation mechanisms of the India-Asia collision: Earth and Planetary Science Letters, v. 86, p. 240-252.

Copeland, P., Harrison, T. M., Pan Y., Kidd, W. S. F., Roden, M., & Zhang Y., 1995, Thermal evolution of the Gangdese batholith, southern Tibet: A history of episodic unroofing: Tectonics, v. 14, p. 223–236,

Coulon, C., Maluski, H., 13011 inger, C., & Wang, S., 1986, Mesozoic and Cenozoic volcanic rocks from central and southern Tibet: $^{39}Ar/^{40}Ar$ dating, metrological characteristics and geodynamical significance: Earth and Planetary Science Letters, v. 79, p. 281–302.

Cross, T, A., & Pilger, R, H., Jr., 1982, Controls of subduction geometry, location of magmatic arcs and tectonics of arc and back-arc regions: Geol. Sot. Am, Bull., v. 93, p. 545–562.

Derbyshire, E., Shi, Y. F., Li, J. J., Zheng, 6. X., Li, S. J., & Wang, J. T., 1991, Quaternary glaciation of Tibet—the geological evidence: Quaternary Science Reviews, v. 10, p. 485-510.

Dewey, J. F., & Burke, K. C. A., 1973, Tibetan, Variscan and Precambrian reactivation: Products of continental collision: Journal of Geology, v. 81, p. 683-692.

Dewey, J. F., Shackelton, R. M., Chengfa, C., & Yiyin, S., 1988, The tectonic evolution of the Tibetan Plateau: Royal Society of London Philosophical Transactions, ser. A, v. 327, p. 379-413.

England, P., & Houseman, G., 1986, Finite strain calculations of continental deformation, 2, Comparison with the India-Asia collision zone: Jour. Geophys. Res., v. 91, p. 3664–3676.

England, P., & Houseman, G., 1988, The mechanics of the Tibetan Plateau: Royal Society of London Philosophical Transactions, ser. A, v. 326, p. 301–320.

England, P., & Houseman, G., 1989, Extension during continental convergence, with application to the Tibetan plateau: Jour. Geophys. Res., v. 94, p, 17,561-17,579.

England, P., & Searle, M., 1986, The Cretaceous-Tertiary deformation of the Lhasa block and its implications for crustal thickening in Tibet: Tectonics, v. 5, p. 1–14.

England, P,, 1995, Late-Tertiary-Quaternary dynamics of the Tibetan plateau: Mitteilungen aus dem Geologischen Institut der Eidgenössischen Technischen Hochschule und der Universität Zurich, Neue Folge, Nr. 298.

Fielding, E. J., [sacks, B, L., Barazangi, M., & Duncan, C. C., 1994, How flat is Tibet?, Geology, v. 22, no. 2, p. 163–167,

Foster, D. A., Gleadow, A, J. W., & Mortimer, G., 1994, Rapid Pliocene exhumation in the Karakoram (Pakistan), revealed by fission-track thermochronology of the K2 gneiss: Geology, v, 22, p. 19-22.

Gansser, A., 1964, The Geology of the Himalayas, Interscience, New York, 289 p.

Gansser, A., 1981, The geodynamic history of the Himalaya: in Zagros, Hindu Kush, Himalaya-Geodynamic Evolution, edited by H.K. Gupta and F.M. Delany, AGU, Washington D. C., p. 111-121.

Gaudemer, Y., Jaupart, C., & Tapponnier, P., 1988, Thermal control on post-orogenic extension in collision belts: Earth Planet. Sci. Lett., v. 89, p. 48–62.

George, M., Reddy, S., & Harris, N., 1995, Isotopic constraints on the cooling history of the Nanga Parbat-Haramosh Massif and Kohistan arc, western Himalaya: Tectonics, v. 14, p. 237–252.

Guillot, S., Hodges, K. V., LeFort, P., & Pécher, A., 1994, New constraints on the age of the Manaslu leucogranite: Evidence for episodic tectonic denudation in the central Himalayas: Geology, v. 22, p. 559–562.

Harris, N. B. W., Ronghua, W., Lewis, C. L., & Chengwei, J., 1988, Plutonic rocks of the 1985 Tibet Geotraverse, Lhasa to Golmud: Royal Society of London Philosophical Transactions, ser. A, v. 327, p. 145-168,

Harrison, T. M., Copeland, P., Kidd, W, S. F., & Yin, A., 1992a, Raising Tibet: Science, v. 255, p. 1663-1670.

Harrison, T. M., Chen, W., Leloup, P. H., Ryerson, F. J., & Tapponnier, P., 1992b, An early Miocene transition in deformation regime within the Red River fault zone, Yunnan, and its significance for Indo-Asian tectonics: Jour. Geophys. Res., v. 97, p. 7159-7182.

Heim, A., & Gansser, A., 1939, Central Himalaya, geological observations of the Swiss expedition 1936: Mere. Sot. Helv. Sci. Nat., v. 73/1, p. 1-245.

Hennig, A., 1915, Zur Petrographie und Geologie von Sudwest Tibet: in Southern Tibet, Kung Boktryekeriet, P. A. Stockholm, Norstedt, v. 5, 220p.

Him, A., 1988, Features of the crust-mantle structure of the Himalayas-Tibet: A comparison with seismic traverses of the Alpine, Pyrenean and Variscan erogenic belts: Royal Society of London Philosophical Transactions, ser. A, v. 326, p. 17--32.

Holt, W. E., & Wallace, T. C., 1990, Crustal thickness and upper mantle velocities in the Tibetan Plateau region from the inversion of regional P_{nl} waveforms: Evidence for a thick upper mantle lid beneath southern Tibet: Jour. Geophys. Res., v. 95, p. 12,499-12,525.

Houseman, G. A., McKenzie, D. P., & Molnar, P., 1981, Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergent belts: Jour. Geophys. Res., v. 86, p. 61 15–61 32.

Hubbard, M. S., & Harrison, T. M., 1989, $^{40}Ar/^{39}Ar$ age constraints on deformation and metamorphism in the MCT Zone and Tibetan Slab, eastern Nepal Himalaya: Tectonics, v. 8, p. 865--880.

Isacks, B. L., 1988, Uplift of the Central Andean Plateau and bending of the Bolivian Orocline: Jour. Geophys. Res., v. 93, p. 3211-3231.

Isacks, B. L., 1992, 'Long-term' land surface processes: Erosion, tectonics and climate history in mountain belts, in Mather, P. M., cd., Terra-1 Understanding the terrestrial environment: London, Taylor and Francis, p. 21–36.

Jin, Y., McNutt, M. K., & Zhu, Y., 1994, Evidence from gravity and topography data for folding of Tibet: Nature, v. 371, p. 669–674.

Johnson, M. R. W., 1994, Volume balance of erosional loss and sediment deposition related to Himalayan uplifts: Jour. Geol. Sot. London, v. 151, p. 217–220.

Jordan, T. E., Isacks, B. L., Allmendinger, R. W., Brewer, J.A, Ramos, V. A., & Ando, C. J., 1983, Andean tectonics related to geometry of subducted Nazca plate: Geol. Sot, Amer. Bull., v. 94, p. 341–361.

Kay, S. M., Maksaev, V., Moscoso, R., Mpodozis, C., & Nasi, C., Probing the evolving Andean lithosphere–-mid-late Tertiary magmatism in Chile (29°–30030'S) over the modern zone of subhorizontal subduction: Jour. Geophys. Res., v. 92, p. 6173–6189.

Kidd, W. S. F., & Molnar, P., 1988, Quaternary and active faulting observed on the 1985 Academica Sinica-Royal Society Geotraverse of Tibet: Royal Society of London Philosophical Transactions, ser. A, v. 327, p. 337–363.

Kidd, W. S. F., Pan Yusheng, Chang Chengfa, Coward, M. P., Dewey, J. F., Gansser, A., Molnar, P., Shackleton, R. M., & Sun Yiyin, 1988, Geological mapping of the 1985 Chinese–British Tibetan (Xizang-Qinghai) plateau geotraverse route: Royal Society of London Philosophical Transactions, ser. A, v. 327, p. 287–305.

Korzoun, V. I., editor-in-chief, 1977, Atlas of world water balance: Paris, UNESCO Press.

Kutzbach, J. E., Prell, W. L., & Ruddiman, W. M., 1993, Sensitivity of Eurasian climate to surface uplift of the Tibetan Plateau: Journal of Geology, v. 101, p. 1 77–1 90.

Le Pichon, X., Fournier, M., & Jolivet, L., 1992, Kinematics, topography, shortening, and extrusion in the India-Eurasia collision: Tectonics, v. 11, p. 1085–1 098.

Leeder, M. R., Smith, A. B., & Yin J., 1988, Sedimentology, palaeoecology and palaeoenvironmental evolution of the 1985 Lhasa to Golmud Geotraverse: Royal Society of London Philosophical Transactions, ser. A, v. 327, p. 1 07–1 44.

Leloup, P. H., Harrison, T. M., Ryerson, F. J., Chen W., Q. Li, Tapponnier, P., & Lacassin, R., 1993, Structural, metrological and thermal evolution of a Tertiary ductile strike-slip shear zone, Diancang Shari, Yunnan: Journal of Geophysical Research, v. 98, p. 6715-6743.

Lerch, F. J., Nerem, R. S., Putney, B. H., Felsentreger, T. L., Sanchez, B.V, *et al.*, 1994, A geopotential model from satellite tracking, altimeter, and surface gravity data - GEM-T3: Journal of Geophysical Research, v. 99, p. 2815–2839.

Lewis, C.L.E, 1990, Thermal history of the Kunlun Batholith, N. Tibet, and implications for uplift of the Tibetan Plateau: Nucl. Tracks Radiat. Meas., v. 17, p. 301-307.

Littledale, St. G. R., 1896, A journey across Tibet, from north to south, and west to Ladak: Geogr. Jour., v. 7, p. 453--483.

Loughridge, M. S., 1986, Relief map of the earth's surface: EOS Trans. AGU, v. 67, p. 121.

Masek, J., Isacks, B.L., Gubbels, T. L., & Fielding, E. J., 1994a, Erosion and tectonics at the margins of continental plateaus: Journal of Geophysical Research, v. 99, no. B7, p. 13,941–13,956.

Masek, J. G., Isacks, B.L., Fielding, E. J., & Browaeys, J., 1994b, Rift-flank uplift in Tibet: Evidence for a viscous lower crust: Tectonics, v. 13, no. 2, p. 659–667.

McKenna, L. W., & Walker, J. D., 1990, Geochemistry of crustally derived leucocratic igneous rocks from the Ulugh Muztagh area, Northern Tibet, and their implications for the formation of the Tibetan Plateau: Journal of Geophysical Research, v. 95, p. 21,483-21,502.

Molnar, P., & Tapponnier, P., 1975, Cenozoic tectonics of Asia: Effects of a continental collision: Science, v. 189, p. 419-426.

Molnar, P., & Tapponnier, P., 1978, Active tectonics of Tibet: Journal of Geophysical Research, v. 83, p. 5361–5375.

Molnar, P., 1988, A review of geophysical constraints on the deep structure of the Tibetan Plateau, the Himalaya and the Karakoram, and their tectonic implications: Royal Society of London Philosophical Transactions, ser. A, V. 326, p. 33-88.

Molnar, P., Burchfiel, B. C., Zhao Z., Liang K., Wang S., & Huang M., 1987, Geologic evolution of northern Tibet: results of an expedition to Ulugh Muztagh: Science, v. 235, p. 299-305.

Molnar, P., England, P., & Martinod, J., 1993, Mantle dynamics, uplift of the Tibetan Plateau, and the Indian monsoon: Reviews of Geophysics, v. 31, no. 4, p. 357–396.

Ni, J., & York, J, E., 1978, Late Cenozoic tectonics of the Tibetan Plateau: Journal of Geophysical Research, v. 83, p. 5377-5384.

Norin, E,, 1946, Geological reconnaissance in western Tibet, Reports from the scientific expedition to the northwestern provinces of China under the leadership of Dr. Sven Hedin: Publication 29, part 3, Geology 7, Tryckeri Aktiebolaget,Thule, Stockholm.

Pan, Y., & Kidd, W, S, F., 1992, Nyainqentanglha shear zone: A late Miocene extensional detachment in the southern Tibetan Plateau: Geology, v. 20, p. 775–778.

Pan, Y., Copeland, P., Roden, M, R., Kidd, W. S. F., & Harrison, T, M., 1993, Thermal and unroofing history of the Lhasa area, southern Tibet—Evidence from apatite fission track thermochronology: Nucl. Tracks Radiat. Meas., v. 21, p. 543-554.

Peltzer, G., & Tapponnier, P., 1988, Formation and the evolution of strike-slip faults, rifts, and basins during the India-Asia collision: An experimental approach: Journal of Geophysical Research, v. 93, p. 15,085–15,1 17.

Powell, C. M., & Conaghan, P. J., 197’3, Plate tectonics and the Himalayas: Earth and Planetary Science Letters, v. 20, p. 1–12.

Ratschbacher, L., Frisch, W., Liu, G., & Chen, C., 1994, Distributed deformation in southern and western Tibet during and after the India-Asia collision: Journal of Geophysical Research, v. 99, p. 19,917-19,945.

Raymo, M. E., & Ruddiman, W. F., 1992, Tectonic forcing of late Cenozoic climate: Nature, v. 359, p. 117–122.

Rothery, D. A., & Drury, S. A., 1984, The neotectonics of the Tibetan Plateau: Tectonics, v. 3, p, 19–26.

Ruddiman, W. F., & Kutzbach, J. E., 1989, Forcing of late Cenozoic northern hemisphere climate by plateau uplift in southern Asia and the American West: Journal of Geophysical Research, v. 94, no. D15, p. 18,409-18,427.

Ruddiman, W. F., Prell, W.L. & Raymo, M. E., 1989, Late Cenozoic uplift in southern Asia and the American West: rationale for general circulation modeling experiments: Journal of Geophysical Research, v. 94, no D15, p. 18,379–18,391.

Schärer, U., Copeland, P., Harrison, T. M., & Searle, M. P., 1990, Age, cooling history, and origin of post-collisional leucogranites in the Karakoram Batholith; A multi-system isotope study: Jour. of Geology, v. 98, p. 233–251.

Searle, M., & 10 others, 1987, The closing of the Tethys and the tectonics of the Himalaya: Geological Society of America Bulletin, v. 98, p. 678–701.

Searle, M., 1986, Structural evolution and sequence of thrusting in the High Himalaya, Tibetan-Tethys and Indus suture zones of Zanskar and Ladakh, Western Himalaya: Jour. Structural Geology, v. 8, p. 923–936.

Searle, M., 1995, The rise and fall of Tibet: Nature, v. 374, p. 17–18.

Searle, M. P., Waters, D, J., Rex, D. C., & Wilson, R. N., 1992, Pressure, temperature and time constraints on Himalayan metamorphism from eastern Kashmir and western Zanskar: Jour. Geol. Sot. London, v. 149, p. 753–773.

Shackleton, R. M., & Chang C., 1988, Cainozoic uplift and deformation of the Tibetan Plateau: the geomorphological evidence: Royal Society of London Philosophical Transactions, ser. A, v. 327, p. 365--378.

Smalley, R., Jr., & Isacks, B. L., 1987, A high-resolution local network study of the Nazca plate Wadati-Benioff zone under western Argentina: Journal of Geophysical Research, v. 92, p. 13,903–1 3,912.

Sorkhabi, R. B., & Stump, E., 1993, Rise of the Himalaya: a geochrologic approach: GSA Today, v. 3, p. 85–92.

Sorkhabi, R. B., 1993, Time-temperature pathways of Himalayan and Trans-Himalayan crystalline rocks: a comparison of fission-track ages: Nucl. Tracks Radiat. Meas., v. 21, p. 535--542.

Stüwe, K., White, L., & Brown, R., 1994, The influence of eroding topography on steady-state isotherms. Application to fission track analysis: Earth and Planetary Science Letters, v. 124, p. 63--74.

Tang B., Liu Y., Zhang L., Zhou W., & Wang Q., 1981, Isostatic gravity anomalies in the central portion of the Himalayas: in Geological and Ecological Studies of the Qinghai-Xizang Plateau, Beijing, Science Press, p. 683–689.

Turner, S., Hawkesworth, C., Liu, J., Rogers, N., Kelley, S., & van Calsteren, P., 1993, Timing of Tibetan uplift constrained by analysis of volcanic rocks: Nature, v. 364, p. 50-54.

Wager, L. R., 1937, The Arun River drainage pattern and the rise of the Himalaya: Geographical Journal, v. 89, p. 239-250.

Wissmann, H. von, 1959, Die heutige Vergletscherung und Schneergrenze in Hochasien: Mainz, Akademie Wissenschaften und der Literatur, Abhandlungen tier Mathematisch-Naturwissenschaftlichen Klasse, p. 5–307.

Yin, A., Harrison, T. M., Ryerson, F. J., Chen W., Kidd, W, S. F., & Copeland, P., 1994, Tertiary structural evolution of the Gangdese thrust system, southeastern Tibet: Journal of Geophysical Research, v. 99, p, 18,1 75–18,201.

Zeitler, P. K., 1985, Cooling history of the NW Himalaya, Pakistan: Tectonics, v. 4, p. 127-151.

Zhao, W., & Morgan, W. J., 1985, Uplift of Tibetan Plateau: Tectonics, v, 4, p. 359–369.

Zhou W., Yang Z., Zhu H., & Wu L., 1981, Characteristics of the gravity field and the crustal structure in the eastern and central regions of the Xizang plateau: in Geological and Ecological Studies of the Qinghai-Xizang Plateau, Beijing, Science Press, p. 673–682.

Zhou X., Cao Y., Zhu M., Xia D., & Qian D., 1986, Plate tectonic-lithofacies map of Xizang (Tibet), China(1:1 ,500,000 scale): Beijing, Geological Publishing House.

Figures

Figure 1, Overview map of Tibet showing major features over smoothed topog-raph y *(darker gray shading >4500 m and lighter gray >3000 m). Thin lines* enclose zone of internal drainage of Tibet, used for hypsometry of Figure 5. *Lines* A–A' and B–B' mark locations of profiles shown in Figure 4. Lambert conformal conic projection is used for Figures 1, 3, and 4.

Figure 2. Profiles of four active erogenic belts, taken perpendicular to strike. Each profile represents the mean elevation from 100-km-wide swaths, Profiles are approximately aligned at their steepest mountain front and are filled with shades and patterns. Tibet profile *(light gray)* runs roughly S–N (left to right), Altiplano profile (medium *gray)* runs NE–SW, New Zealand profile *(dark gray)* is NW–SE, and Taiwan profile *(wavy lines)* runs E–W,

Figure 3. Shaded relief of Central Asia topography at -500 m resolution. Topography was illuminated from northeast.

Figure 4. Slope map calculated from full resolution topography. Slopes are measured within -250 m windows and sampled at -1 km resolution. White areas are missing data. Gray scale bar shows slope values.

Figure 5. Profiles of Tibetan Plateau, taken along 100-km-wide swaths. A: North-trending profile A–A' across widest part near longitude 85°E. B: Longitudinal profile B-B' along Tibet length (see Fig. 1 for locations). Each profile shows maximum, minimum (shading), and mean elevations (thick solid lines) of the topography within 5 x 100 km segments, along with topographic relief (thin lines). Thick dashed lines indicate annual precipitation; scale is on right.

Figure 6. Hypsometry of internally drained area of central Tibet (see Fig. 1 for location), Differential (solid) and cumulative (dashed) hypsometry have the same vertical (elevation) axis and different horizontal axes (top and bottom). Thick solid line indicates mean elevation.

Figure 7. Tibet isostatic uplift history scenario. F or five approximate times (1 00, 45, 25, 14, and O Ma), the estimated lithospheric structure is shown as a column with the mantle lithosphere, crust below sea level, and crust above sea level shaded according to the key. The lithosphere scale is to the left. For the same times, the elevation of the surface is also shown as black squares, vertically exaggerated according to the scale at the right.

Figure 8. Annual average precipitation for Tibet and surrounding areas adapted from Climatic Atlas of Asia (1981), Key shows shading patterns for precipitation. Outline of Tibet topography shown as lines for geographic reference.

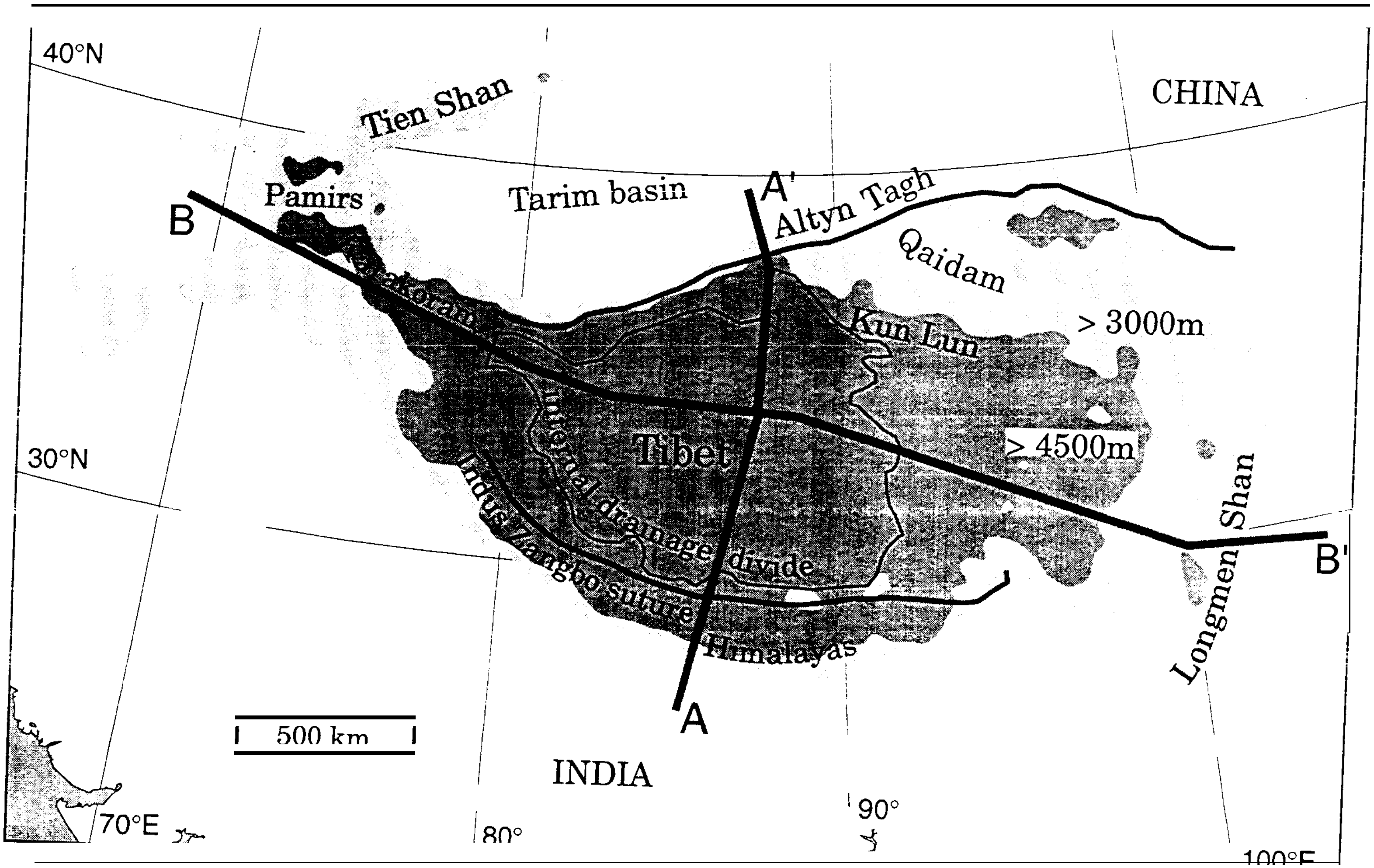
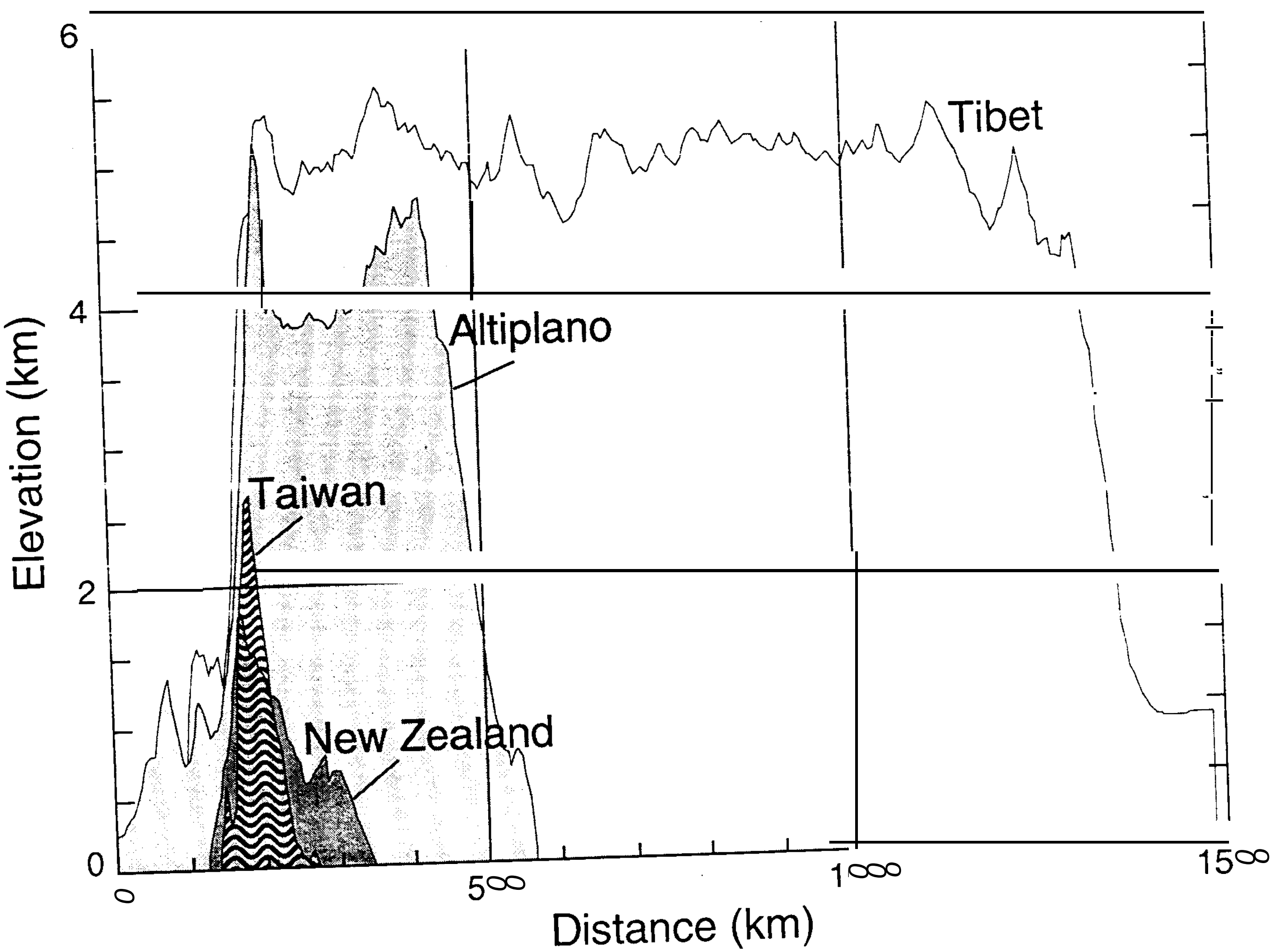

40°N
30°N
Tien Shan
CHINA
Pamirs
Tarim basin
A'
Altyn Tagh
Qaidam
B
Kun Lun
> 3000m
Tibet
> 4500m
Internal drainage divide
Indus Zangbo suture
Himalayas
Longmen Shan
B'
A
500 km
INDIA
70°E
80°
90°
100°E

Figure 1



Tibet
Altiplano
Taiwan
New Zealand
Elevation (km)
6
4
2
0
Distance (km)
0
500
1000
1500

Figure 2

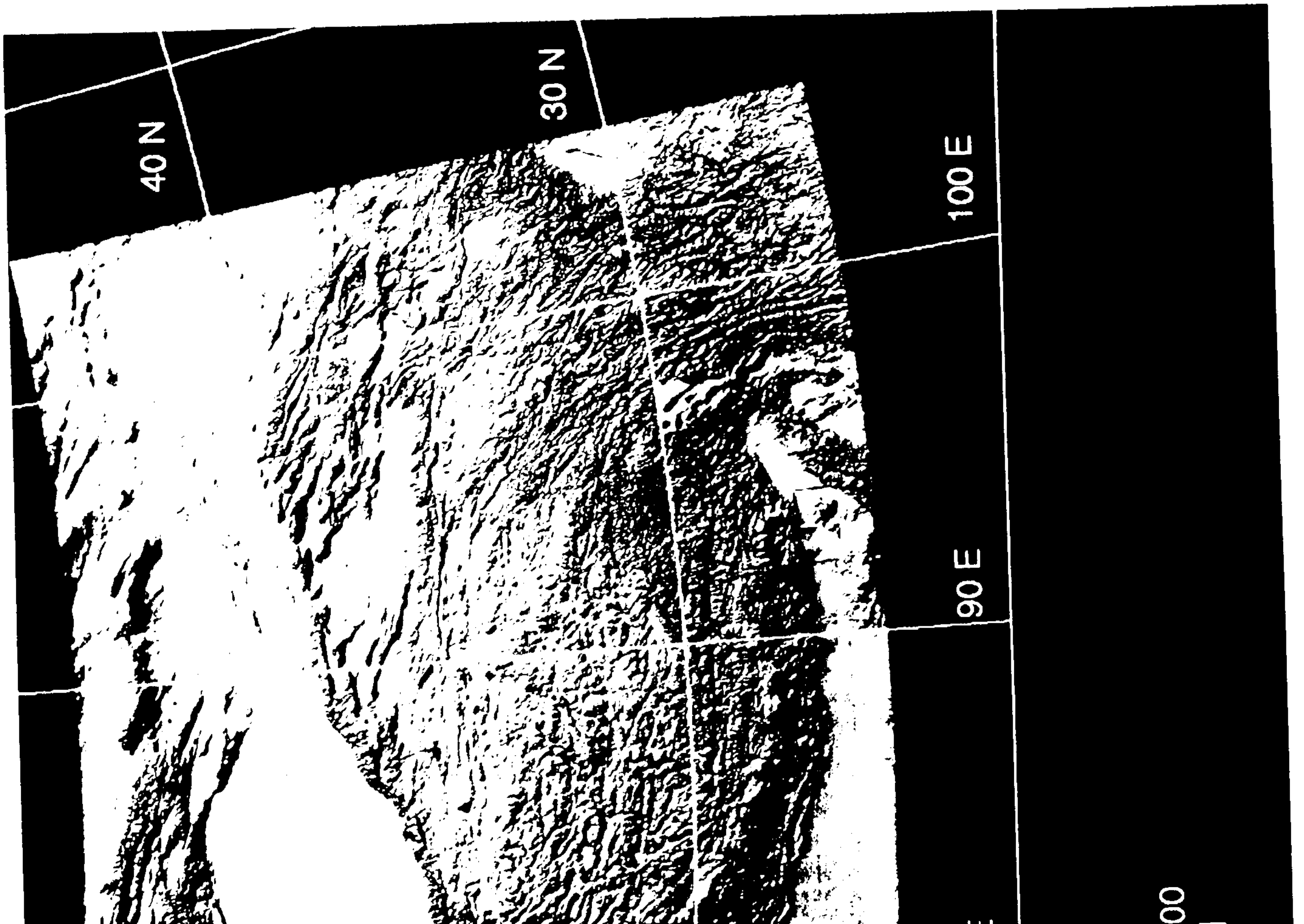


Figure 3

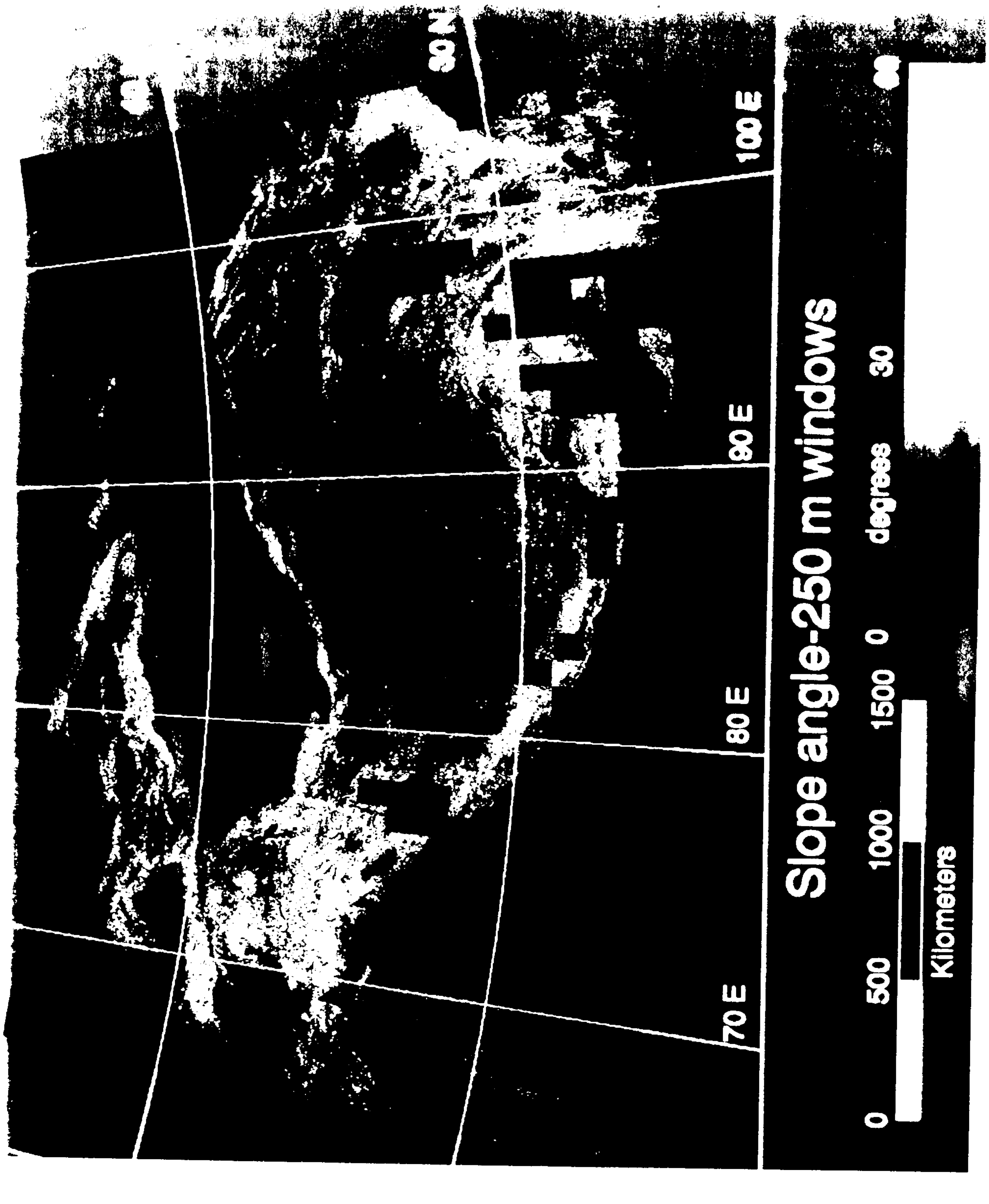

Slope angle-250 m windows
degrees
0
30
0
500
1000
1500
Kilometers
70 E
80 E
90 E
100 E
30 N

Figure 4

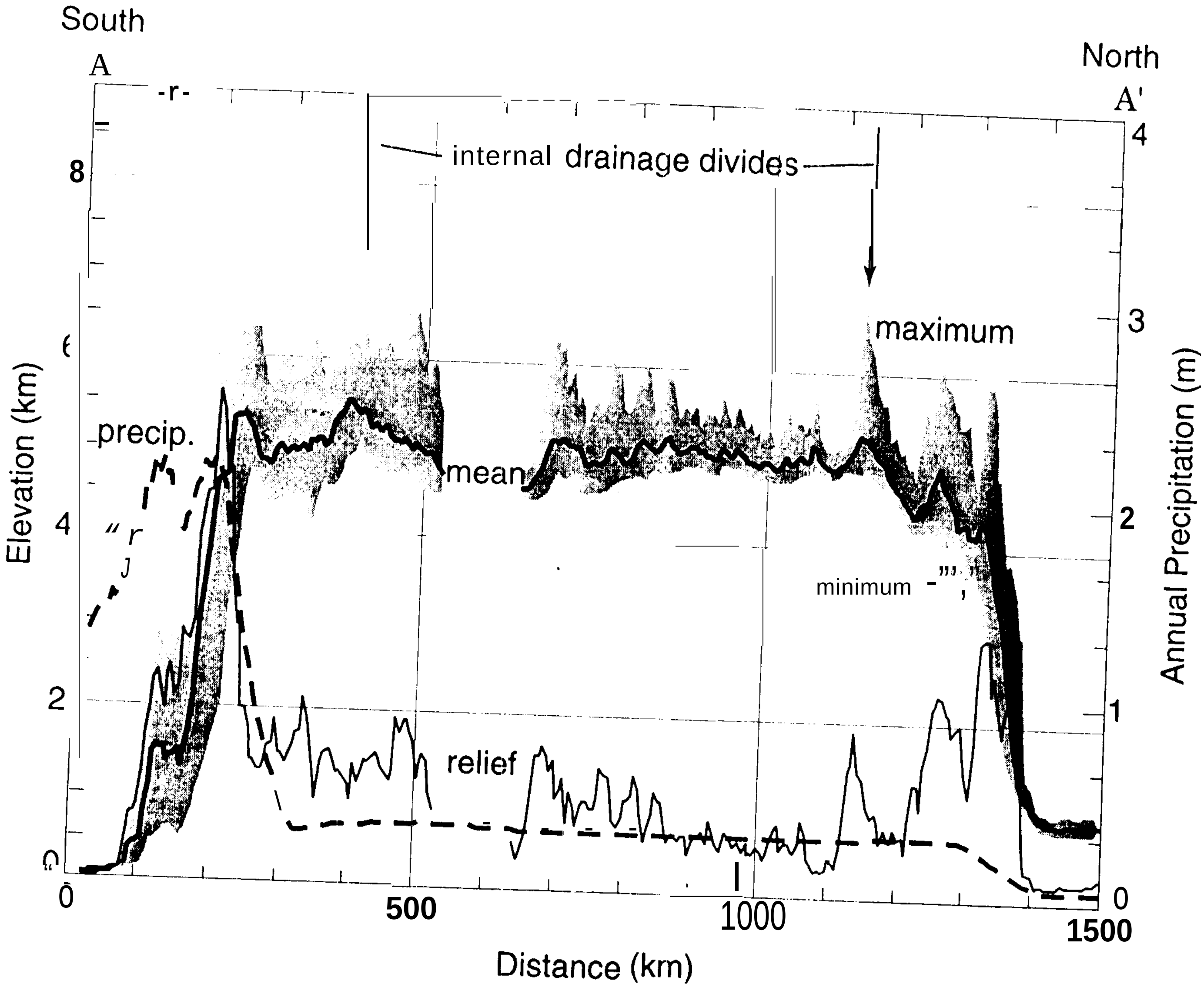

South
A
North
A'
internal drainage divides
maximum
precip.
mean
minimum
relief
Elevation (km)
Annual Precipitation (m)
Distance (km)
0
500
1000
1500
8
4
2
0
4
3
2
1
0

Figure 5(a)

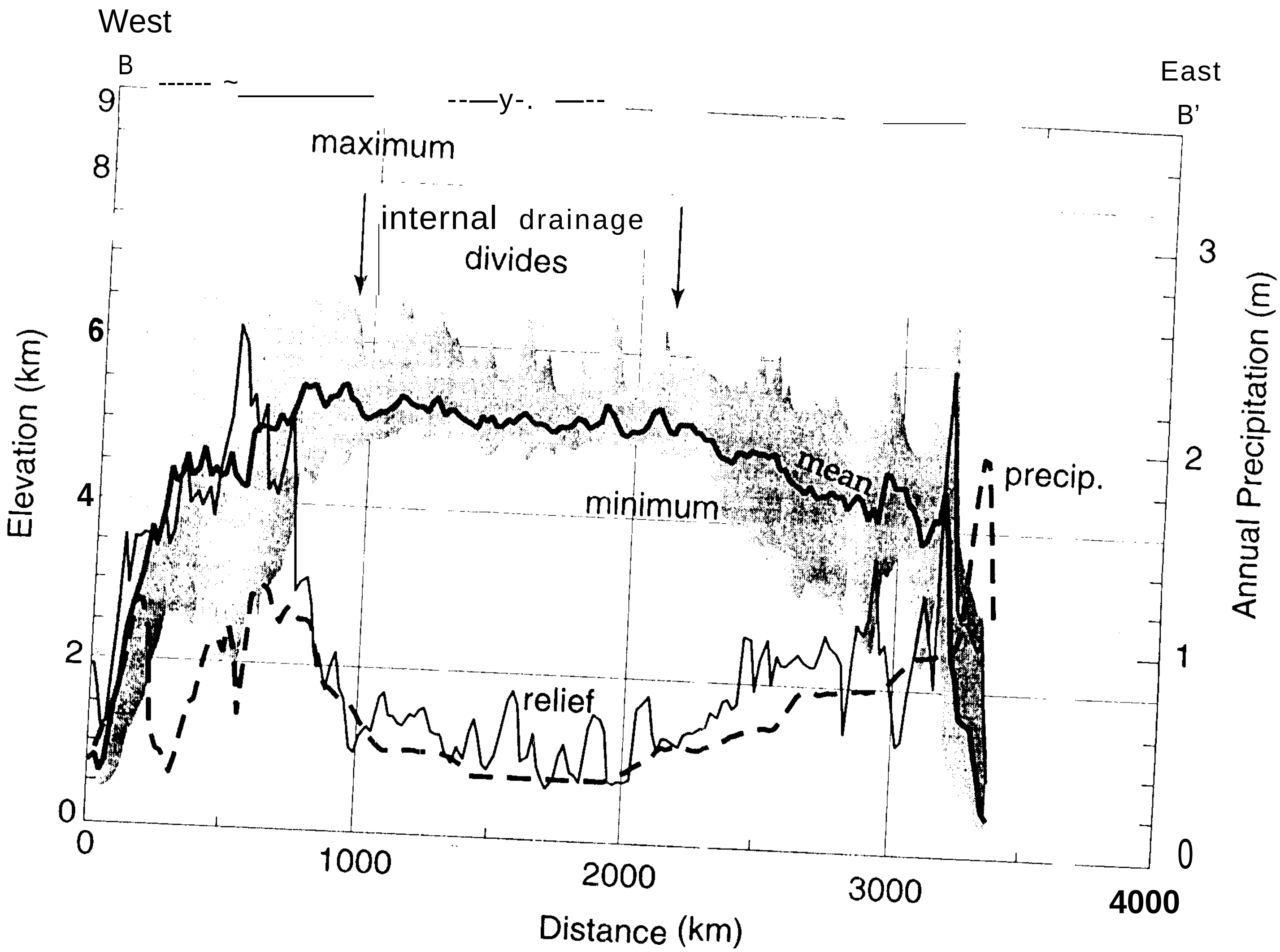

West
B
East
B'
maximum
internal drainage
divides
mean
precip.
minimum
relief
Elevation (km)
Annual Precipitation (m)
Distance (km)
0
2
4
6
8
9
1
3
1000
2000
3000
4000

Figure 5(b)

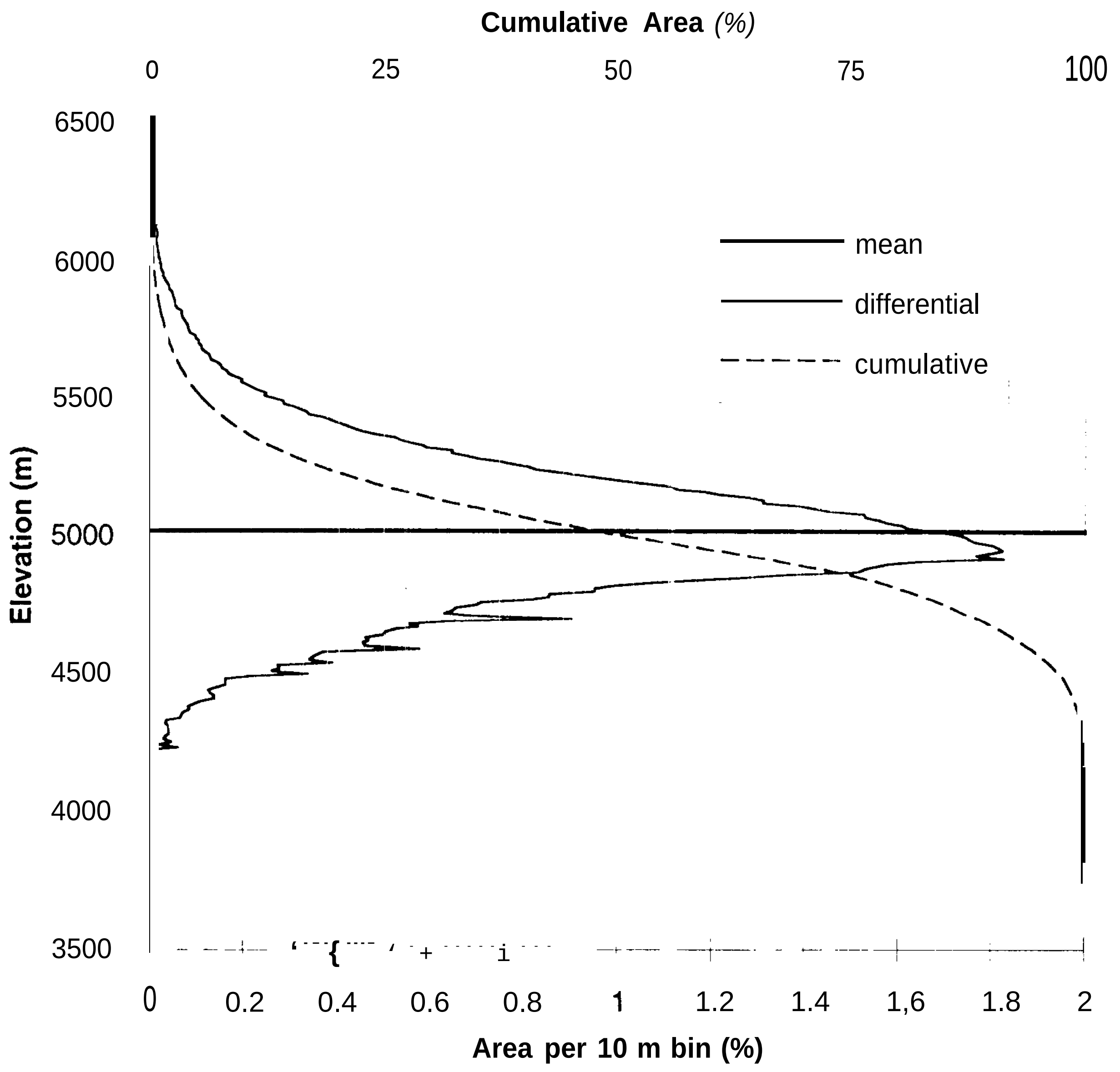
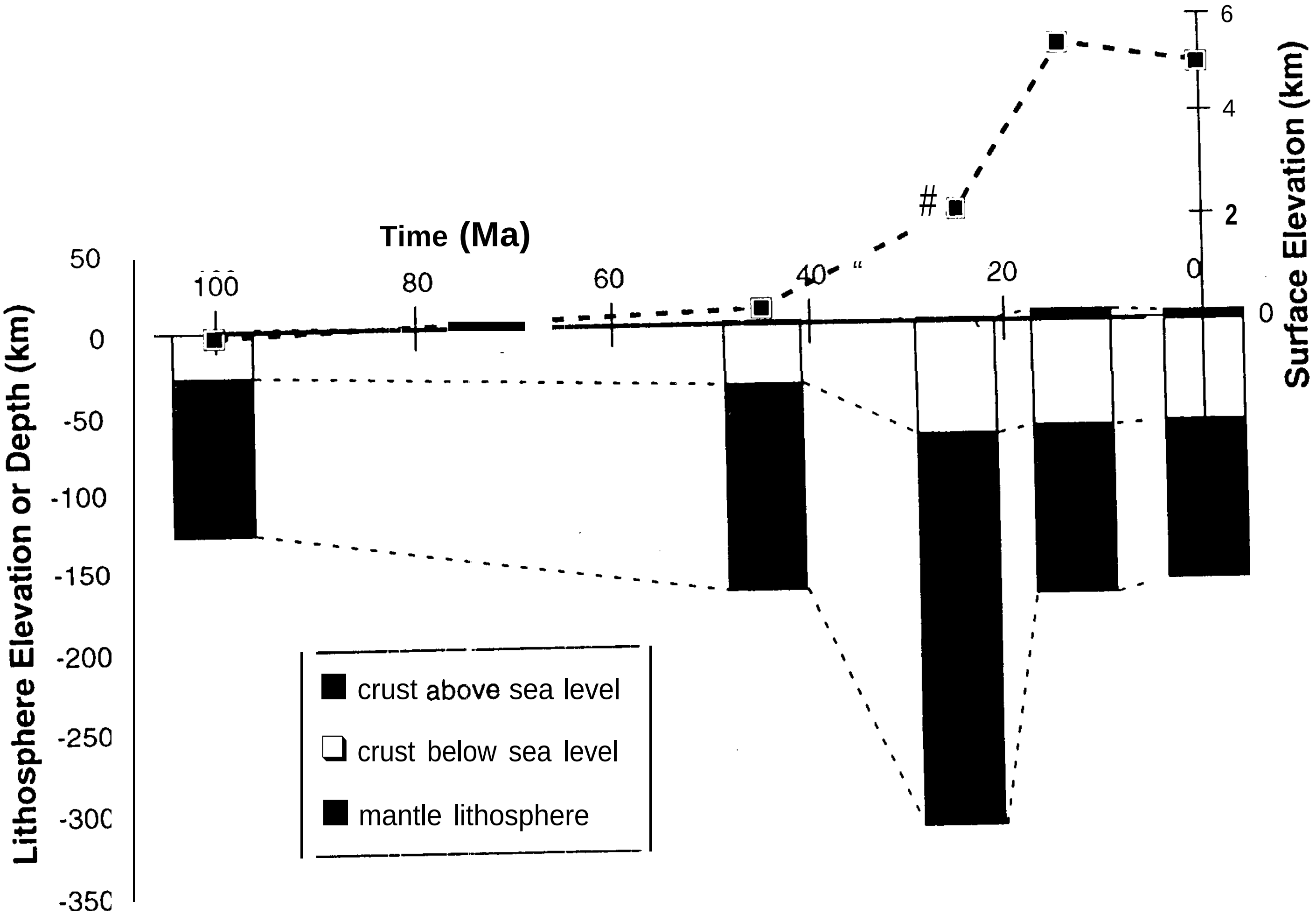

Cumulative Area (%)
0
25
50
75
100
6500
6000
5500
5000
4500
4000
3500
Elevation (m)
mean
differential
cumulative
0
0.2
0.4
0.6
0.8
1
1.2
1.4
1,6
1.8
2
Area per 10 m bin (%)

Figure 6



Time (Ma)
100
80
60
40
20
0
50
0
-50
-100
-150
-200
-250
-300
-350
Lithosphere Elevation or Depth (km)
6
4
2
0
Surface Elevation (km)
crust above sea level
crust below sea level
mantle lithosphere

Figure 7

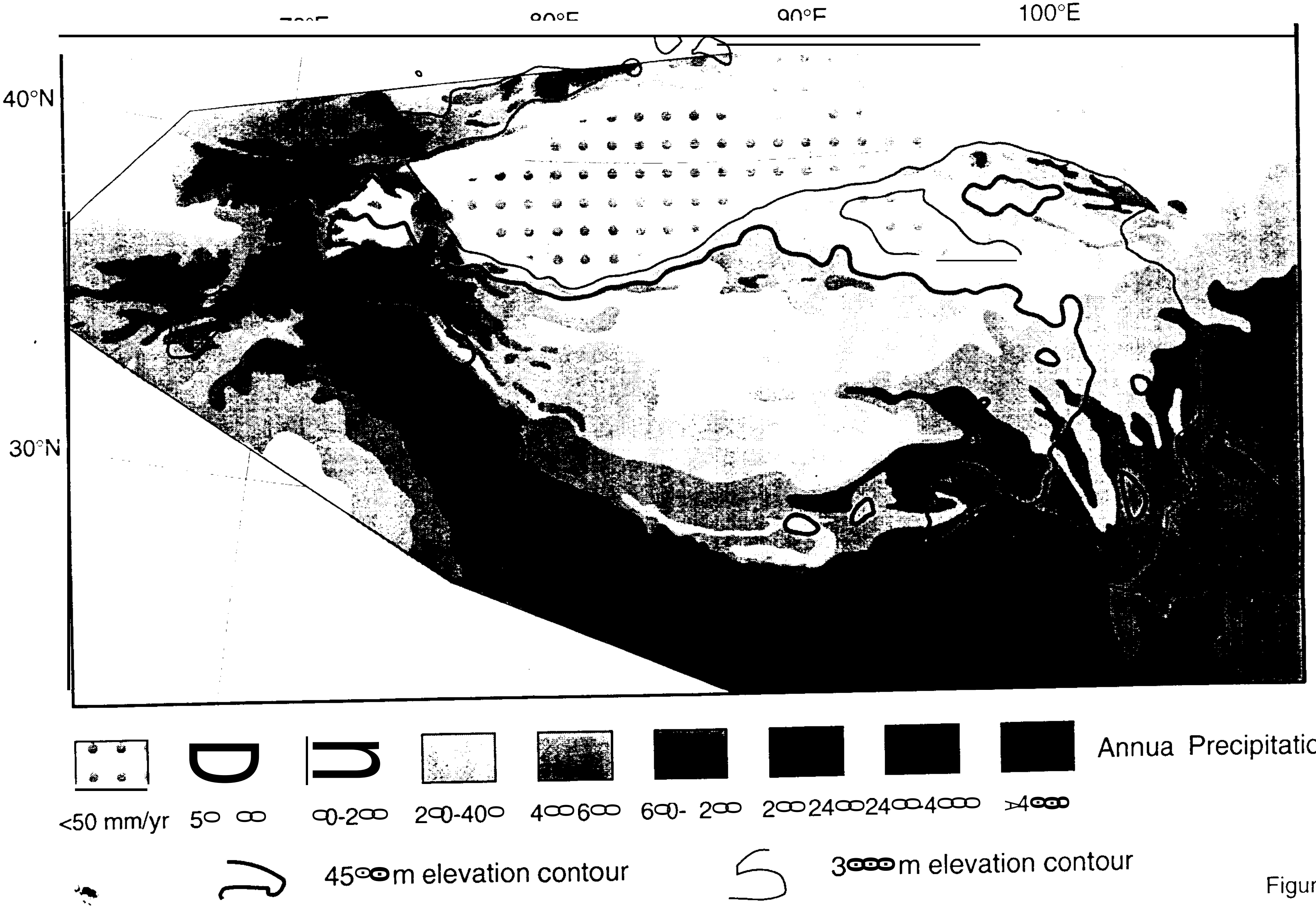

100°E
40°N
30°N
<50 mm/yr
Annua Precipitatio
4500m elevation contour
3000m elevation contour

Figure 8